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NEW ENGLAND INTERCOLLEGIATE
GEOLOGIC CONFERENCE
GUIDEBOOK

for

THE GREENVILLE - MILLINOCKET REGIONS, NORTH CENTRAL, MAINE

EDITORS

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75th Annual Meeting

October 7, 8 and 9, 1983

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NEW ENGLAND INTERCOLLEGIATE GEOLOGIC CONFERENCE

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for field trips in

The Greenville - Millinocket Regions, North Central, Maine

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CONFERENCE ORGANIZATION

CO-CHAIRMAN

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TABLE OF CONTENTS

page

Conference organization ii

Forward iv

Introduction to the geology of North-Central Maine, D.W. Caldwell and Lindley S. Hanson 1

Trip A-1: **Glacial traverse across the southern side of an ice cap,** Thomas V. Lowell 11

Trip B-1: **The Hurricane Mountain Formation Melange and unconfomably overlying Lower to Middle Ordovician volcanics, Brassua Lake and Moosehead Lake Quadrangles,** Gary M. Boone 31

Trip B-2: **Geology of rocks along the East Branch Pleasant River, White Cap Range, Central Maine,** Lindley S. Hanson 45

Trip B-3: **Features of Upper Silurian and Lower Devonian sedimentary rock in the Caucomgomoc Lake Area, Northwestern Maine,** Steven G. Pollock 63

Trip B-4: **The timing of alpine glaciation of Mt. Katahdin,** D.W. Caldwell and P.T. Davis 79

Trip B-5: **The Northwest Boundary Fault of the Boundary Mountain Anticlinorium,** David S. Westerman 87

Trip C-1: **The petrology of the central portion of the Moxie Pluton,** Rudolph Hon and Jaye Schulman 103

Trip C-2: **Stratigraphy, metamorphism and geomorphology in the Greenville-Rockwood Area, Maine** , D.W. Caldwell and Lindley S. Hanson 117

Trip C-3: **Stratigraphy and sedimentation in Silurian flysch East of Millinocket, Maine,** David C. Roy, Nancy A. Demarest, and Maureen J. Hills 123

Trip C-4: **The Seboomook Lake Area became ice free; but how?** Thomas V. Lowell and Kristine Crossen 147

Trip C-5: **Ophiolite and melange terrane, Caucomgomoc Lake Area, Northwestern Maine,** Steven G. Pollock 175

Trip D-1: Repeat of C-1

Trip D-2: Repeat of C-2

Trip D-3: **Multiple till localities near Bigham and New Sharon,** D.W. Caldwell and Thomas Weddle 191

Forward

The 75th annual meeting is being conducted at the edge of and partly within the great north woods of Maine. Access to these woods is mostly by roads constructed within the past 30 years by wood-products companies, notably the Great Northern Paper Company, Scott Paper Company, St. Regis, and Diamond International. In addition to owning the cutting rights to these woods, these companies also own the mineral rights to these lands. For this reason they have been happy to have geologic studies conducted in this region and have been most helpful to those of us working on their lands.

The Rangley Lakes region to the southwest of this conference area was presented th during the 62nd Annual Meeting, Gary M. Boone, editor. The geology of the Presque Isle region and neighboring New Brunswick to the northeast was covered during the 72nd Annual Meeting, David C. Roy and Richard S. Naylor, editors. The guidebook presented here bridges the gap between these two important works. Since the earlier conferences Gary Boone has continued his studies toward the northeast and Dave Roy toward the southwest. We are pleased that both will be leading field trips during this meeting.

At the same time that this guidebook is being prepared, the Maine Geological Survey is putting the finishing touches on two new geologic maps of the State, one of the bedrock geology and the other of the surficial geology. Many of the guidebook authors have been involved in the preparation of these maps and much of the geology that will be shown in this conference was mapped in the past few years by these authors for these new maps. In this way, the Maine Geological Survey has played a major role in this conference, as it has in previous conferences during the past 23 years.

New England Intercollegiate Geologic Conference(NEIGC)

The NEIGC was began in 1903, when William Morris Davis led an informal field trip to the Connecticut Valley of Massachusetts. As in that first meeting, the sole purpose of the NEIGC has always been to conduct field trips in areas of recent geologic mapping. The 75th Annual Meeting is the 8th meeting in the State of Maine since 1960. The high frequency of trips in Maine is in recent years is a product of the support that the modern Maine Geological Survey and its State Geologists, John R. Rand, Robert G. Doyle, and Walter A. Anderson have given to geologic studies within the State.

INTRODUCTION TO THE GEOLOGY OF NORTH-CENTRAL MAINE

by

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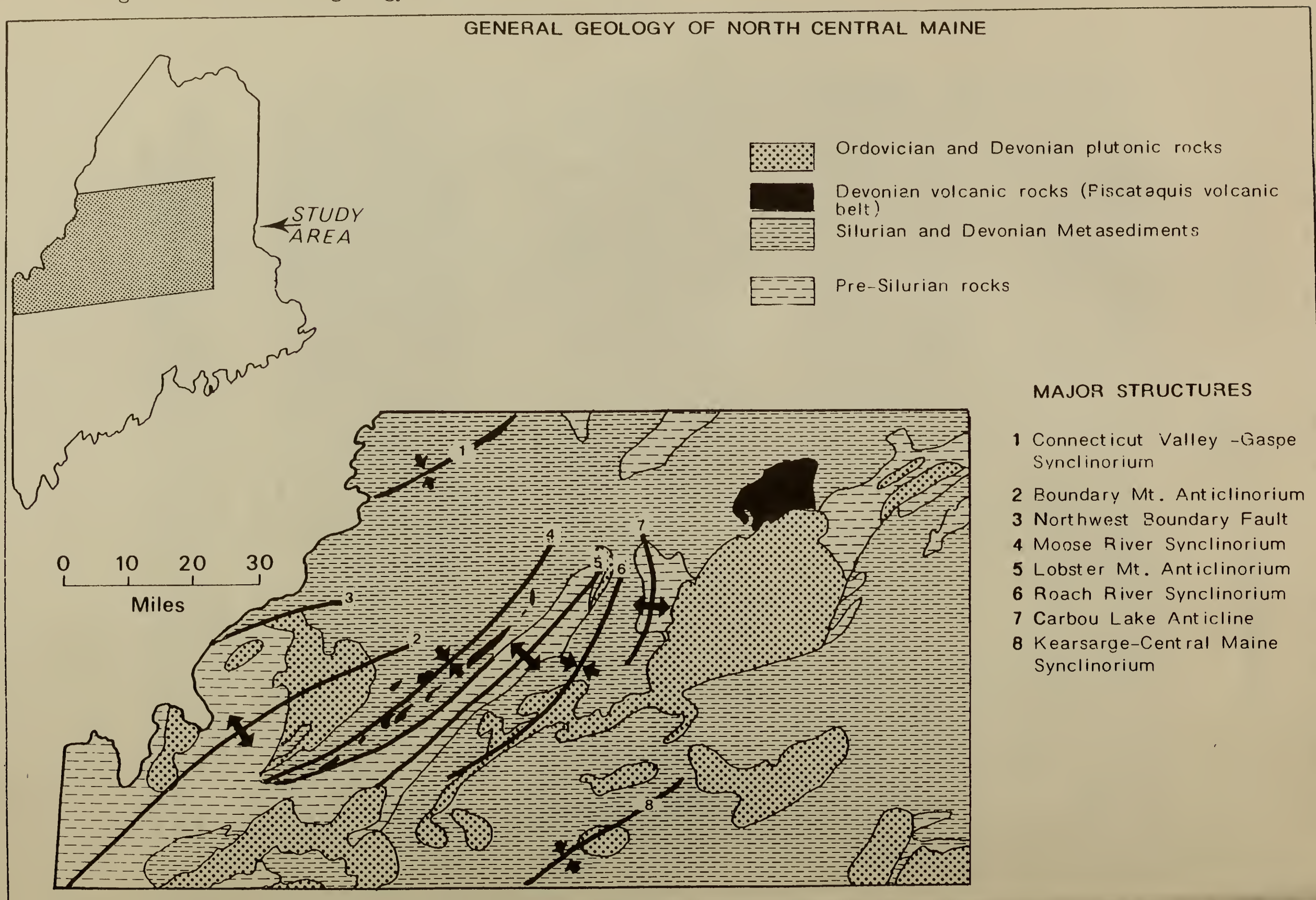
Together the fieldtrips of this conference traverse several major SW-NE structural belts; from the Boundary Mountain Anticlinorium in the west to the Kearsarge-Central Maine Synclinorium (Lyons and others, 1982) in the east. These belts encompass a wide variety of rocks which span over a billion years from Precambrian to Lower Devonian times. The fieldtrips of this conference investigate and illustrate many of the tectonic, stratigraphic, and geomorphic features which characterize the geology of this region. For a summary of the stratigraphies and principal formations of this refer to figure 2 and table I.

The western portion of the conference area is occupied by the Boundary Mountain Anticlinorium (fig. I), a major structural belt which is the northeastern extension of the Bronson Hill Anticlinorium to the south. The anticlinal core is occupied in part by the Chain Lakes Massif, the oldest exposed Precambrian basement in New England, dated at 1.5 billion years (Naylor and Others, 1973). Also exposed are Cambrian and Ordovician volcanic, metasedimentary, and plutonic rocks. The Boundary Mountain Anticlinorium plunges NNE where it eventually disappears beneath a blanket of Upper Silurian and Lower Devonian metasediments. The northwestern limb of the anticlinorium is truncated by the Northwest Boundary Fault (see Westerman, trip B-5), a major thrust which can be traced southwestward into New Hampshire.

East of the Boundary Mountain Anticlinorium lie the Moose River and Roach River Synclinoria, separated by the complimentary Lobster Mountain Anticlinorium (Boucot, 1961: Boucot and Heath, 1969). The Moose River Synclinorium has a SW-NE doubly plunging axis and is asymmetrical with a steeply dipping southeast limb. Several minor folds, reflecting similar geometries, lie within the synclinorium and expose rocks of the Lower Devonian Tarratine and Tomhegan formations. The Lower Devonian Seboomook and a variety of Silurian rocks are exposed along the southeastern margin of the synclinorium. The above-mentioned formations and similar structural geometries are also displayed in the Roach River Synclinorium to the east.

The intervening Lobster Mountain Anticlinorium encompasses metasedimentary and volcanic rocks of both Cambrian and Ordovician age (See Boone, trip B-I). These include; the Hurricane Mountain Formation, an accretionary melange containing exotic blocks of basalt and gabbro;

Figure 1. Generalized geology of North-central Maine.



	MOOSE RIVER-ROACH SYNCLINORIA	CAUCOMGOMOC LAKE REGION	NORTHWEST LIMB OF KEARSARGE-CENTRAL MAINE SYNCLINORIUM		PLUTONIC ROCKS
LOWER DEVONIAN	Tomhegan Kineo Volcanic Member	Seboomook	Seboomook(?)		Katahdin Pluton Moxie Pluton
	Tarratine Seboomook				
	Carrabassett		Carrabassett(?)		
	?	?			
SILURIAN	Capens	Allagash Lake	Madrid(?)	Lawler Ridge	
	undifferentiated Silurian		Smalls Falls	unnamed rusty unit	
			Perry Mt.	Allsbury	
ϵ-O	Lobster Mountain Volcanics	?			Attean Pluton (Jackman area)
CAMBRIAN(?)	Dead River				
	Hurricane Mountain	Hurd Mountain			
	Jim Pond	Caucomgomoc Lake Loone Stream			

Figure. 2. Generalized stratigraphic column of conference area.

TABLE 1. DESCRIPTIONS OF PRINCIPAL ROCK UNITS IN AREA OF CONFERENCE. MORE COMPLETE DESCRIPTION IN MOENCH AND OTHERS (1982)

Stratified Rocks

Devonian Rocks	Formation
Dto	Tomhegan Formation (Boucot, 1961; Boucot and Heath, 1969). These rocks are restricted in occurrence to the central region of the Moose River Synclinorium. The formation consists of dark metasandstone, dark tuffaceous metasandstone, slate, metasiltstone and quartzite. The formation contains abundant fossils, mainly brachiopods. The type locality is in Tomhegan Township, Brassua Lake Quadrangle.
Dtokm	Kineo member of the Tomhegan Formation (Boucot, 1961; Boucot and Heath, 1969). The Kineo member consists of a variety of rhyolitic rocks, including felsite, tuff, and volcanic breccia and conglomerate. The Kineo occurs at the edge of the Moose River Synclinorium and is correlated with similar volcanic rocks in Big Spencer Mountain, Sorburge Mountain in the Harrington Lake Quadrangle and with the Traveler Rhyolite (Hon, 1980). The formation is named for exposures in Kineo Mountain, Moosehead Lake Quadrangle.
Dt	Tarratine Formation (Boucot, 1961; Boucot and Heath, 1969). The Tarratine consists of dark metasandstone, slate and metasiltstone and is highly fossilferous. It is roughly equivalent in age with the Matagamon sandstone of Rankin (1965) and apparently represents shallower water deposition than was present during the deposition of the underlying slates and turbidites. The Tarratine is exposed along the margins of the Moose River Synclinorium. The type locality is on Misery Ridge in the Brassua Lake Quadrangle.
Ds	Seboomook Formation (Boucot, 1961; Boucot and Heath, 1969). The Seboomook is a cyclically bedded well-graded metasandstone and slate. The sandy layers weather to a chalky white and commonly show cross laminations.

The Seboomook is widespread in northwestern Maine and occurs in a broad belt between Greenville and the Katahdin Pluton. South of the Moxie Pluton is restricted within the area of this conference to the cores of synclines within the underlying Carrabassett Formation.

The type locality of the Seboomook is at Seboomook Lake in the Seboomook Lake Quadrangle. The Seboomook is considered to be partly equivalent to the Littleton Formation of New Hampshire and Gile Mountain of Vermont.

Dc — Carrabassett Formation (Boone, 1973). The Carrabassett is divided into an upper cyclicly bedded but poorly graded metasandstone that is similar to the Seboomook and a lower massive pelitic member. In earlier maps (e.g., Doyle, 1967), the rocks now mapped as Carrabassett were mapped as Seboomook. The slate quarries between Monson and Brownville are in the Carrabassett.

The Carrabassett is partly equivalent to the Littleton New Hampshire. The type locality is in the headwaters area of the Carrabassett River in the Little Bigelow Mountain Quadrangle.

Silurian Rocks

Sm — Madrid Formation (Osberg and others, 1968). The Madrid is primarily a thickly bedded pale gray to greenish-gray feldspalthic metasandstone (Moench and others, 1982). In the type locality in the village of Madrid the formation contains pods and lenses of calc-silicate minerals.

The Lawler Ridge (Roy and others, this volume) is an equivalent rock northeast of Millocket.

Su — Silurian rocks, undifferentiated.

Ordovician Rocks

Ov — Ordovician volcanic rocks. This is a grab bag term for disparate volcanic rocks including the Kennebec volcanics and the Lobster Mountain volcanics, both exposed in the Lobster Mountain Anticlinorium (Boucot and Heath, 1969).

Cambro-Ordovician(?) Rocks

€Od — Dead River Formation (Boone, 1973). Metasiltstone and phyllite with minor feldspathic metasandstone. Typically the Dead River is greenish to greenish grey with a pearly phyllitic luster. The phyllite has three well developed cleavages, causing it to break off in sharp edged rhombic fragments.

The Dead River is exposed within the Lobster Mountain Anticlinorium. The type locality is near the Dead River in the northern part of the Little Bigelow Mountain Quadrangle.

Cambrian(?) Rocks

€h — Hurricane Mountain Formation (Boudette and Boone, 1976). A sedimentary melange consisting of a rusty weathering metasiltstone incorporating exotic blocks of granite, quartzite, amphibolite, and other types of rocks.

The type locality is on Hurricane Mountain in the Pierce Pond quadrangle. Within the conference area the unit is exposed in the Lobster Mountain Anticlinorium and is exposed within a fault-bounded block in the Coucomgomac Lake area where Pollack (this volume) has found a similar unit which he named the Hurd Mountain Formation.

€jp — Jim Pond Formation (Boudette, 1982). The Jim Pond consists of basalt flows and pillowed basalts with minor metagreywacke. Pllock (this volume) has named apparently equivalent rocks the Coucomgomac Lake Formation. These rocks are part of an ophiolite suite named the Boil Mountain complex by Boudette (1982). The formation is named for exposures near Jim Pond in the Jim Pond quadrangle. Within the conference area, the Jim Pond is exposed in the Lobster Mountain anticlinorium and in the Coucomgomac Lake area as described above.

Intrusive Igneous Rocks

Devonian Rocks

Dkp — Katahdin Pluton (Griscom, 1976; Hon, 1980). The Katahdin Pluton is an oval shaped laccolith with long axis trending NE-SW. The major rock within the pluton is the Katahdin granite, and the principal rock type within the granite is the Doubletop facies, a massive biotite granite of medium texture.

Most of the Doubletop Facies underlies lowland areas, except where it is capped by the very resistant, fine-grained Summit Facies (Hon, 1980). The Summit Facies has a tough, interlocking texture that resists erosion. Where the Summit Facies is removed, the Doubletop and other intermediate facies are quickly eroded.

Hon (1980) interprets the Katahdin as an S-type granite and follows other workers who believe that the Katahdin was part of the magma chamber for the Traveler Rhyolite.

Dmp — Moxie Pluton (Espenshade and Boudette, 1964; Espenshade, 1972). The Moxie Pluton is an elongate, irregular-shaped mafic pluton extending for about 50 miles (80km) from Moxie Mountain on the southwest to the southwestern end of the Katahdin Pluton on the northeast. It ranges in width from as little as a mile to more than 9 miles (15km) in several bulbous emanations spaced along the length of the pluton.

The Moxie is a highly differentiated mafic pluton, ranging from quartz diorite and diorite near the Katahdin Pluton, through troctolite and norite in the Greenville area, to dunite on the southwest end of the pluton at Moxie Mountain. Hon (1980), this volume, interprets this distribution of rock types to mean that the Moxie was originally formed as a vertically differentiated body that was then tilted upward toward the southwest and subsequently eroded to a common surface. Thus in moving along the pluton from the northeast to southwest one goes from the mafic-poor rocks of the upper part of the pluton to the mafic-rich cummulate rocks of the lower part of the pluton.

The Moxie breaks down rapidly when exposed to chemical and mechanical weathering and thus underlies low areas.

Ordovician Rocks

Oap — Attean Pluton (Albee and Boudette, 1972). The Attean is a pink or grey, two-feldspar porphyritic granite. It commonly is altered with the area of the pluton to unicite, containing pink feldspars and green epidote and quartz. The Attean is extensively jointed and fractured. These surfaces showing a rusty stain.

The Attean occurs in northwestern Maine in the Jackman region. It is abundant in both the Attean and Skinner quadrangles and is apparently named for exposures along Attean Lake in the quadrangle of the same name.

the green, phyllitic Dead River Formation, a Cambro-Ordovician(?) flysch; and the volcanic and volcanoclastic rocks of the Lobster Mountain Volcanics. Similar rocks are exposed in an isolated fenster through Upper Silurian and Lower Devonian metasediments in the Caucomgomoc Lake area to the north (Pollock, trip C-5).

The above fold belts are separated from the Kearsarge-Central Maine synclinorium by the Greenville Plutonic belt which includes the granitic Katahdin Pluton and the highly differentiated, mafic Moxie Pluton (Hon and Schulman, trip C-1). These rocks comprise the plutonic core of a volcanic arc formed by subduction during the Acadian Orogeny (Rankin, 1968; Hon, 1983). The Piscataquis volcanic belt (fig. 1), which includes the Kineo (lower member of the Tomhegan Formation) and Traveler rhyolites, is what remains of the thick volcanic piles formed during this time. These volcanics comprise some of the youngest rocks in the region and have been preserved from post-Acadian uplift and erosion only where down-folded in the troughs of synclines, such as the Kineo in the Moose River Synclinorium, or down-faulted, as in the Traveler Caldera (Rankin, 1968).

Contact aeroles of granofels are common where late Acadian plutons intruded pelitic rocks or the Roach River and Kersearge-Central Maine Synclinoria(Hanson, trip B-2, and Caldwell and Hanson, trip C-2). The width of the aeroles vary and is generally governed by the temperature of the pluton during implacement and the complexity of the contact.

The Eastern margin of the conference area is occupied by the Kearsarge-Central Maine Synclinorium, a belt dominated by Silurian and Devonian flysh which extends southwest through central Maine and New Hampshire (Roy, Trip C-3, and Hanson Trip, B-2).

The major SW-NE fold belts in this region are the product of the Acadian Orogeny. Folding was accompanied by low-grade regional metamorphism and the development of vertical to nearly vertical penetrative cleavage in pelitic rocks. Locally rocks were contact metamorphosed by the intrusion of Late Acadian plutons.

The present landscape of the region is the result of upward tilting of the orogen to the southeast followed by several kilometers of erosion. Ridges and valleys characterize the differential erosion within fold belts and basins of deeply weathered, coarse-grained plutonic rock are surrounded by irregular mountains of granofels.

Glaciation, which is responsible for only a few meters of erosion, has not greatly altered the topography of this region. Glacial deposits average only a few meters in thickness, but locally may exceed several tens of meters. The only record of glacial deposition in Maine during the Pleistocene dates from the Wisconsin Stage, with a possible early Wisconsin till and soil exposed at New Sharon (Caldwell and Weddle, Trip D-3).

The Late Wisconsin ice sheet advanced to its terminal moraine on the continental shelf about 17,000 years ago. The ice sheet subsequently retreated

to near the present Maine coast about 13,500 years ago (Stuiver and Borns, 1975). For the next 1,000 years the ice sheet was marine based, with the rate of ice retreat determined by the depth of water and tidal conditions rather than by climatic factors.

The incursion of the sea occurred as the ice retreated from the coast. The greatest transgression of the sea into Maine occurred in the large river valleys; to near Millinocket in the Penobscot Valley; north of Brownville Junction in the Pleasant River; near Bigham in the Kennebec Valley; and near Dixville in the Androscoggin Valley. By the time the ice sheet retreated to near the present Maine Coast the Champlain Sea separated the Laurentide Ice in the St Lawrence Lowlands and an ice cap was stranded over northern Maine. Flow from the northern Maine ice cap was both northwestward into the St. John Valley and southeastward into the headwaters of the Penobscot drainage (Lowell, trip A-1).

There is a continued debate over the timing of Alpine glaciation at Mt. Katahdin (Caldwell and Hanson, 1982; Davis, 1983; Caldwell and Davis, trip B-4). Caldwell believes alpine glaciation followed the maximum Late Wisconsin continental glaciation, while Davis believes there is no evidence for Late Wisconsin alpine glaciation on Mt. Katahdin.

The final melting of the remnant ice cap northwest of Katahdin left small, scattered deposits of sand and gravel, perhaps indicating wholesale stagnation (Lowell, trip C-4).

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GLACIAL TRAVERSE ACROSS THE SOUTHERN SIDE OF AN ICE CAP

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INTRODUCTION

During the last decade considerable debate has developed between workers in southern New England and workers in the Maritime Provinces over glaciation style. Field data from southern New England shows a progressive retreat of the ice front from it's maximum limit on Long Island northward through Connecticut, Rhode Island, and Massachusetts into New Hampshire and Vermont. Furthermore, coastal Maine shows recession of active ice to at least the marine limit (Smith, 1982).

In contrast, the Gaspe, New Brunswick, Cape Breton Island and Nova Scotia all show evidence of interconnected and interacting ice masses. Grant (1977) suggested the maximum extent of these during the late Wisconsin may be the near present coastline. Thus ice recession proceeded from several directions at once.

Debate over these two glaciation models has been active for several years and is not likely to be settled soon. However, these two models imply vastly different ice extents and ice limits, so they are important and require rigorous testing. Similar arguments exist in Newfoundland (see Brookes, 1982 for a interesting review).

One way to view the conflict is that it stemmed from chance placement of political boundaries which later dictated different research approaches: Americans developed one model whereas Canadians support a different model. Thus it may be largely the result of misunderstanding, lack of communication, or lack of data in the interlying areas. The primary purpose of my mapping since 1979 is to reduce the last problem. By selecting an unmapped area central to those suggested contrasting glacial styles, one might find evidence supporting one or the other, or one might find a large gray zone between black and white.

A second development helping our present understanding of the region is the compilation of 1:500,000 scale maps of New Brunswick (Rampton, in press) and Maine (Thompson and others, in press). These maps allow the relationships of moraine distribution, esker distribution and changes in ice-flow indicators to be studied at previously unavailable scales.

As these data are analyzed and studied in a regional context, a better understanding of the regional events should emerge. This trip traverses the area of current research, and should be viewed as an open forum for discussion to look at some old problems in a new light.

ICE-FLOW PATTERNS OF NORTHERN MAINE AND ADJACENT AREAS

The study of ice-flow patterns requires that sufficient data over a wide area be assembled before their meaning can be incorporated into glaciation models. Thus it is necessary to draw on the work of others in addition to presentation of original work. In the following, I will present the ice flow patters deduced from my work in northwestern Maine, add pertinent works from southern Maine and New Brunswick, and suggest a glaciation style somewhat in variance to more classical views of the area. Discussion of this suggestion is requested.

From maps published by LaSalle and others (1977), one easily sees northward ice-flow indicators. However, upon closer examination, one also finds localities where the northward flow lies upon evidence of eastward flow. The eastward flow does not seem to have been influenced by the St. Lawrence Valley. Indeed, Occhietti (Fig. 12, 1982) suggested the origin of this flow as being from the Laurentide Park Highlands during growth of the Nouveau-Quebec Ice Cap.

Lowell (1980a, in press) and Kite and others (1982) reported extensive ice-flow evidence from northwesternmost Maine. The first glacier flow is also directed to the east or east-southeast. Grant (personal communication) suggested a pre-late Wisconsin age for this flow. If so, then the superposed extensive northward flow in Maine, Quebec and New Brunswick must represent the flow origination from an independent ice cap over Maine and the Maritime Provinces that persisted throughout the late Wisconsin; this precludes the influence of Laurentide ice in the region. However, based on striation distribution, lack of weathering between the striation sets, and till studies I suggest that it is of regional importance and occurred during the early phases of the late Wisconsin. Becker (1982) supports this interpretation based on studies of stratigraphic sections along the St. John River.

Moving southward from northernmost Maine the evidence for east/ east-southeast ice-flow becomes stronger. In the region of Ross Lake the evidence of northward flowing ice disappears and only the east/east-southeast indicators remain. Hyland (1981) reported, from north central Maine, that the earliest ice-flow was also to the east-southeast. However, superposed upon these indicators are erosional marks indicating that the subsequent ice-flow shifted to a more southerly direction. In multiple striation localities, the very youngest have a due south

orientation. My reconnaissance checks in the fall of 1982 show this to be a widespread occurrence. The eastward flow is interpreted to be a contemptuous event across all of northwestern Maine. The zone situated between evidence of north and southward flow is called an ice divide. This ice divide (Ross Lake Ice Divide) trends in a NE-SW direction and probably connects to the Quebec Ice Divide (Shilts, 1981).

In contrast, this situation may not be present in the eastern part of Aroostook county, where Genes and others (1981) presented a composite map showing the eastward flow shifting in a areal manner toward the south. Their map reports only rare multiple striation localities. However, Leavitt and Perkins (1935) on the basis of striation data and erratic distribution suggest that two flows crossed this region; one to the southeast and a second more to the south. Which of these contrasting views best explain the situation in eastern Aroostook County is problematic but in the following the more widespread distribution reported in Leavitt and Perkins is used. The ice-flow changes could be spatial, temporal, or both in nature.

The southern extent of localities showing southward ice flow over southeastward ice-flow is not known. However, Konoyer (1979) and Lowell (1980) reported, in the Millinocket-Lincoln area, southern flow over the southeastward indicators. Moreover, this region also shows scattered occurrences of eastward striations. To the east, Rampton and Paradis (1981) reported east-west drumlinoid ridges, eastward striations, and localties showing a southward swing in the Woodstock area of New Brunswick. Further the eastward indicators can be traced into Miramichi Bay. Gauthier (1979) also mapped strong early eastward indicators northeast of Miramich Bay. Thus the eastward flow can thus be traced from the St. Lawrence River near Quebec across northern Maine, and possibly south into central Maine and perhaps east into New Brunswick.

It seems that in regions south of the Quebec/Ross Lake ice Divide the flow chronology shows a dominate southeast ice-flow as indicated by lake alignment, drumlins and flutes that has superposed on it a weaker southward ice-flow. This shift has been reported in several widespread studies; the Kennebec Valley (Smith, 1964), Lincoln (Lowell, 1980b), Mount Desert Island (Lowell, 1980a), Bangor (Brady, 1982), Washington County (Ackert, 1982; Holland, 1981). Rather they can all be related to the same cause requires considerable further work but the hypothesis is advanced here for testing.

The key to this mess is to find a good reason for it. Two reasonable suggestions come to mind: 1) Following Grants (1977) hypothesis that the eastward flow originated in the early phases of the Wisconsin, then the subsequent north and south ice-flows represent the outflow from a independent late Wisconsin age ice mass. One consequence is that most of Maine and the Maritimes remained covered with independent ice during most of the Wisconsin. 2) Following Lowell's (in press) suggestion that all

these features represent late Wisconsin events, then Laurentide ice invaded the region during the early part of the late Wisconsin but gave way to independent ice cap activity when cut off by a ice stream draing northeast in the St. Lawrence Seaway area. In either case, the relative sequence of events is the same, only the ages and causes differ. The last ice-flow moved outward from northcentral Maine and other centers.

I feel that an understanding of ice-flow events are necessary to correctly interpret the extent and influence of Laurentide ice in the Maritime Provinces, Maine and even southern New England. If Laurentide ice did not cross the St. Lawrence River into northern Maine, is the assumption that it flowed into the rest of New England still valid ? Perhaps the model of interconnecting ice masses better represents the late Wisconsin maximum ice configuration. Further, the study of ice-flow events provide information on amount and direction of ice activity just prior to deglaciation.

DEGLACIATION STYLE

Now that the manner of late glacial ice-flow has been left in confusion, it is time to consider the manner of deglaciation. The model common throughout southern New England is that of northward marginal recession. Although the minor details of that recession are still being hotly debated, the large scale view shows progressive northward recession with some measure of ice activity as the ice retreated. Moreover, in coastal Maine, extensive moraine sets demonstrate a complex interaction of retreating ice and rising marine waters (Smith, 1982; Thompson, 1982). However, above the marine limit well defined ice positions are not common (Caldwell, 1975; Kenoyer, 1979). Some small moraines, outwash heads, and repeating eskers segments show that the ice margin temporary paused at numerous locations. As yet, these locations have not been traced with any success from one valley to another.

A important consideration in northern New England is that in addition to an ice margin in the Gulf of Maine another ice margin developed to the east in the Gulf of St. Lawrence and third existed to the north in the St. Lawrence valley. Radiocarbon dates show that marine waters invaded the St. Lawrence Lowland a the same time they rose against the Maine coast. The Bay of Chaleur (just 200 km from eastern Aroostook County) opened somewhat earlier. Therefore, if the ice is receding northward from coastal Maine, eastward from the Gulf of St. Lawrence and southward from the St. Lawrence River, what happens in the middle ?

The Concept of Areal Thinning

In contrast to continuous retreating marginal model, the glacier may have simply stagnated over a wide area and downwasted in place. As the ice thinned, high mountains regions are first exposed as nunataks. With continued downward ice recession (as opposed to recession like a window shade) more and more area becomes exposed. The final ice occupied large lowlands. This deglaciation process is best documented in European studies, i.e. Garns and Bergerson, 1980; Seppala, 1980; and Marcussen, 1977. Ice draining into ice streams into Gulf of Maine, the Gulf of St. Lawrence, and the St. Lawrence River, might allow the highlands to become exposed first.

Borns and Calkin (1977) showed that ice thinned and separated over the Boundary Mountains, meltwater flowed around stagnant ice blocks east of the mountains while an ice margin dammed pro-glacial lakes west of the mountains (Shilts, 1981). A second dipstick showing early emergence is Mt. Katahdin. Although Caldwell and Davis (elsewhere in this volume for the latest round) still debate the amount of cirque activity on the east side of the mountain subsequent to emergence, abundant evidence shows ice filling the valleys around the mountain at about the 700 m level.

Lowell (in press) and Kite and others (1982) presented evidence from northern most Maine showing ice recession from the Notre Dame Mountain southeastward. This ice margin dammed a series of pro-glacial lakes that first drained through and then later northeast around the Notre Dame Mountains.

Considering these data and applying them in an areal thinning model requires that the large scale deglaciation of Maine be more analogous to a thinning pancake, rather than a shrinking dome. The rising sea level south, east, and north of Maine allowed large volumes of ice to be moved, via ice streams, from the interior of the ice mass to the margin without a significant retreat of the margin. Following this model, once the relative marine level dropped, it left a somewhat exhausted ice mass that had a wide spread distribution but had a flat profile that filled the lowlands. Final dissipation of a thick ice mass would not produce deposits with climactic importance. Because the activity on Mt. Katahdin is at most only limited, the region likely lies below the regional snowline after the peak emerged.

Support for this model comes from north-central Maine. In northern parts of this region, the striation data shows a strong, unified northward flow but very little evidence of late local divergence forced by topographic control. Once the ice stopped flowing into the Gulf of Maine, the Gulf of St. Lawrence, and the St. Lawrence valley, it had a flat profile that could not maintain even localized flow.

The deglaciation deposits, on the other hand, show considerable topographic control; extensive valleys choked with

ice disintegration drift and stagnation moraine, lack of extensive or significant gravel deposits, and lack of defined ice positions.

HOW DOES THIS TRIP FIT INTO ALL OF THIS ?

All of this regional speculation is fine, but since we can't view it all in one day, just what are we going to see ? In order to accommodate trip logistics (i.e. the shortest road log) a route from Ashland to Rip Dam seems the best bet (Fig. 1). Unfortunately the trip trends parallel to the Ross Lake Ice Divide but is displaced just to the south of it. Therefore we will not see northward, good examples of southeastward striations, or cross-cutting locations, only the east-southeastward ice-flow indicators.

This field trip area is also located at the northern edge of the gravel zone. That is, very few valleys to the northwest contain appreciable or systematic gravel deposits. Ice separated into various basins and downwasted with the meltwater directed out of one or a succession of outlets. In this case limited meltwater reworked drift and deposited restricted kames, kame terraces, outwash fans, and outwash plains. The nature of two of these basins will be one focus of this trip (Stops 1,2,3,4,6,7 8,9).

Further, this region lies just on the southern end of mapping completed in 1982; thus most deposits and features within the area are not fully understood. Nevertheless, I believe the stops chosen exemplify the large scale deglaciation style sufficiently to warrant their examination; interpretations are subject to considerable improvement. Indeed discussion on this trip may prove the best catalyst.

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LOCATION MAP

FIGURE 1

▲ STOP

0 5 10 km

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ROAD LOG

The following describes a route from Ashland to Rip Dam. The mileages are approximate and could change do to logging operations. Most of the trip is through the North Maine Woods; you must check in and out at Six Mile and Telos Gates. Some fees will be charged for private usage. The suggested stops may be modified to accommodate interests of the group, road changes, new discoveries, and acts of God.

Assembly point is Chris Hotel and Diner, on Route 163 just east of Ashland Maine. Starting time is 8:00 A.M. Stops during the trip will be in the Millinocket Lake, Spider Lake, and Churchill Lake U.S.G.S. 15′ quadrangles.

Mileage

0.0	Leave Chris's parking lot, proceed west (left).
0.2	Turn right (N) onto Main Street at flashing light.
0.6	Turn left (W) at second flashing light.
1.4	Turn left (S) after crossing Aroostook River.
1.6	Pass Ashland Logging Museum.
1.9	Turn right (W) onto dirt road at North Maine Woods sign.
6.6	Keep left (SW).
6.9	Keep left (S) just after passing North Maine Woods gate.
7.3	Cross Greenlaw Stream.
15.8	Keep left (S) to cross Machias River.
29.6	Continue past Chase Brook Rd.
30.7	Pinkham Lumber Company Camp.
30.8	***Stop #1.*** Park well off road just after crossing Mooseleuk. Downstream from the bridge is a 5 m thick exposure of

gravel. The flat top of this deposit is at approximately 215 m.

31.2 Turn left (S) at 4 way intersection.

32.2 ***Stop #2.*** Turn left (E) into Rocky Brook gravel pit.

This active pit shows 5 m of bedded gravel. Dipping beds within suggest water flow toward 342 degrees. Moredver, Hall (1981, oral communication) noted the lack of Munsungan Chert clasts in the pit (the unit outcrops extensively to the north).

The flat topped feature is approximately 20 m above the present stream at 230 m. Faults, with up to 30 cm displacement, are present within the beds.

Observations from a pit within the core of a small esker some 6.5 km to the southwest show flow structures suggesting water flow to 115 degrees. Also noted were abundant Munsungan Chert clasts.

Interpretation: The lack of chert in the Rocky Brook Pit requires that the meltwater and debris were produced south of Norway Bluff. In contrast, the data from the esker pit are compatible with ESE ice flow and subsequent ice recession or downwasting. However, one way to accommodate the data from Rocky Brook is to have separate ice blocks disintegrating in adjacent valleys. Meltwater from one of these blocks could be supplied from Rocky Brook or from the south in the Aroostook valley. The delta may pre-date the Mooseleuk Stream outwash gravels (Stop 1).

33.1 Return to 4 way, continue straight (N).

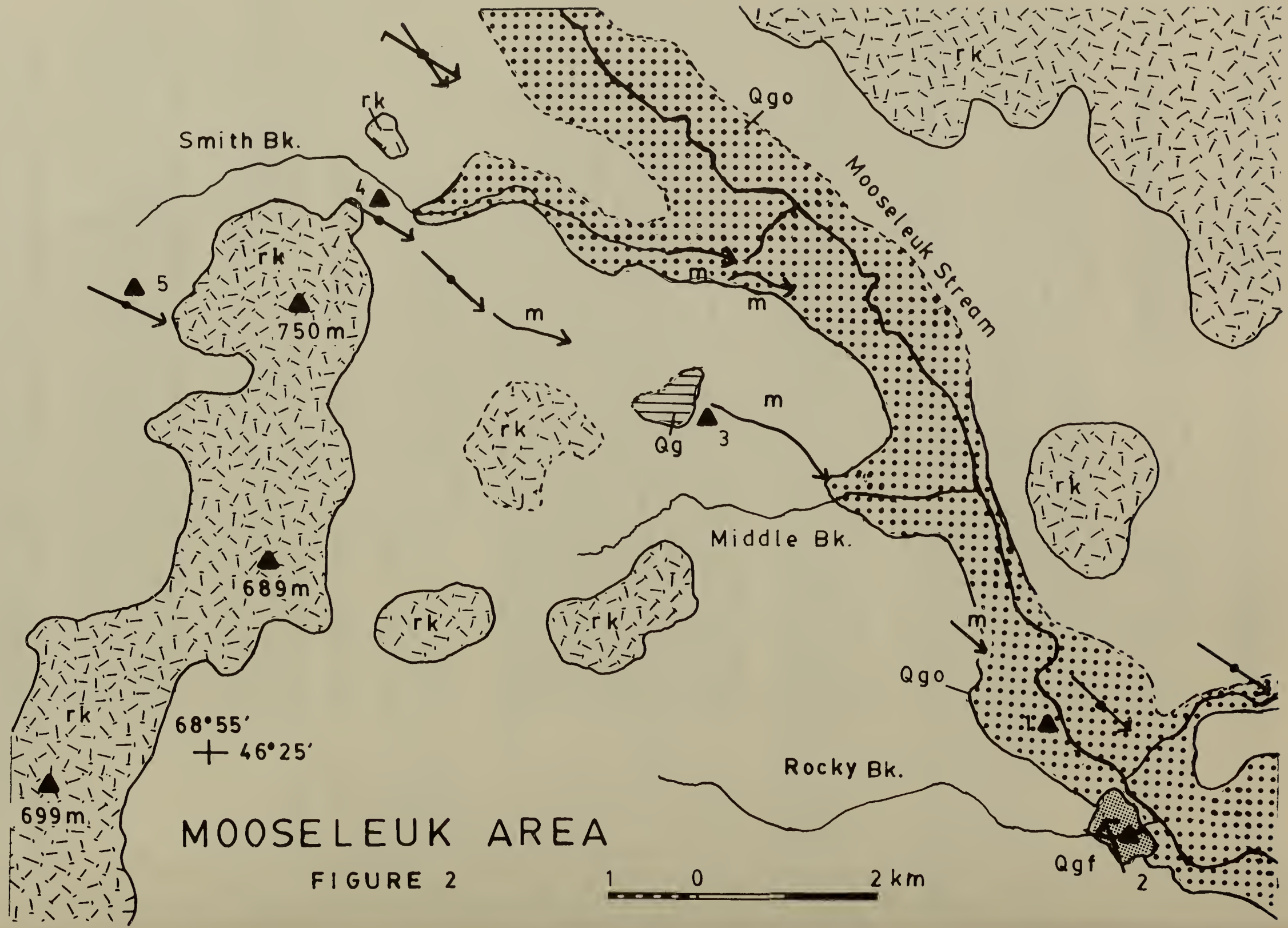
33.7 Keep left (NW).

36.1 ***Stop #3.*** Meltwater channel crosses road.

For the last .2 mi the road has run parallel to a small, 10 m wide, but long (1.5 km) meltwater channel. A small gravel deposit lies near the channel head at 300 m elevation.

Interpretation: This channel probably drained meltwater from a north facing valley across this flat and into the valley of Middle Brook (Fig. 2). This channel may be nearly contemporaneous with other channels to the northeast.

Figure 2. Overleaf. Surficial geology of the Mooseleuk area. m = meltwater channel, Qg = gravel, Qgf = gravel fan, Qgo = outwash, rk = rock. Striation localities = arrow with dot.



MOOSELEUK AREA

FIGURE 2

39.6 ***Stop #4.*** Striated rock and overlaying till.

The drift here displays two facies: the upper 30 cm is fissile and loose whereas the lower (up to 3 m thick) is compact, has a silty matrix supporting 20 per cent stones. Below the drift is banded volcanic bedrock displaying excellent stoss and lee forms and striation with an azimuth of 122 degrees. Below and north of this location we can see the 600 m wide gap through which Smith Brook flows.

Interpretation: The upper drift is probably colluvium derived from the underlying basal till. The striations are aligned with the valley trend but they are also nearly parallel with striations at the next stop. Smith Brook contributed meltwater and gravel into the Mooseleuk Stream from ice held behind (northwest) Mooseleuk Mountain.

41.8 Keep right (N) at Y.

41.9 ***Stop #5.*** Striated outcrop on south road side.

The outcrop south of the road shows a uniform direction of 115 degrees.

Interpretation: At this locality ice had to rise up at least 165 meters to pass over Mooseleuk Mountain. The close agreement of several striation localities (including stop 4) suggest that they were cut under an ice mass capable of ignoring at least 300 meters of relief.

43.0 Forest Service Camp.

47.0 Pass through Pell & Pell Camp.

47.2 Continue past road to south. This road provides access to the Archaeology site on Chase Lake.

47.9 Cross Chase Stream.

48.6 Access Road to Chase Lake. During the lunch stop park in the pit on north side of the road.

The campsite, located on the lake shore provides an excellent place for the black flies to have lunch.

Please note also the poorly exposed gravels in the pit. They are at 265 m elevation and are part of the nearly

continuous gravels that ring Chase Lake at this elevation.

49.6 Turn right (N) just before crossing stream.

50.1 *Stop #6.* Striated outcrop and meltwater channel.

After leaving the main road we climbed above Sewall Deadwater (Fig. 3) to an elevation of 283 m. The outcrop on the roads north side exhibits an 107 degree trend. Stoss and lee features require ice moving east-southeast through this gap. Around the outcrop a small meltwater channel is cut into the valley side.

50.6 Return to main road, turn right (W), and cross stream.

50.7 *Stop #7.* Turn right (N) into Munsungan Brook gravel pit.

This pit, located within a small hill on the valley floor, shows highly variable stratification and undulating contacts. Rare Canadian Shield erratics have been noted here.

Upon leaving pit turn right (W)

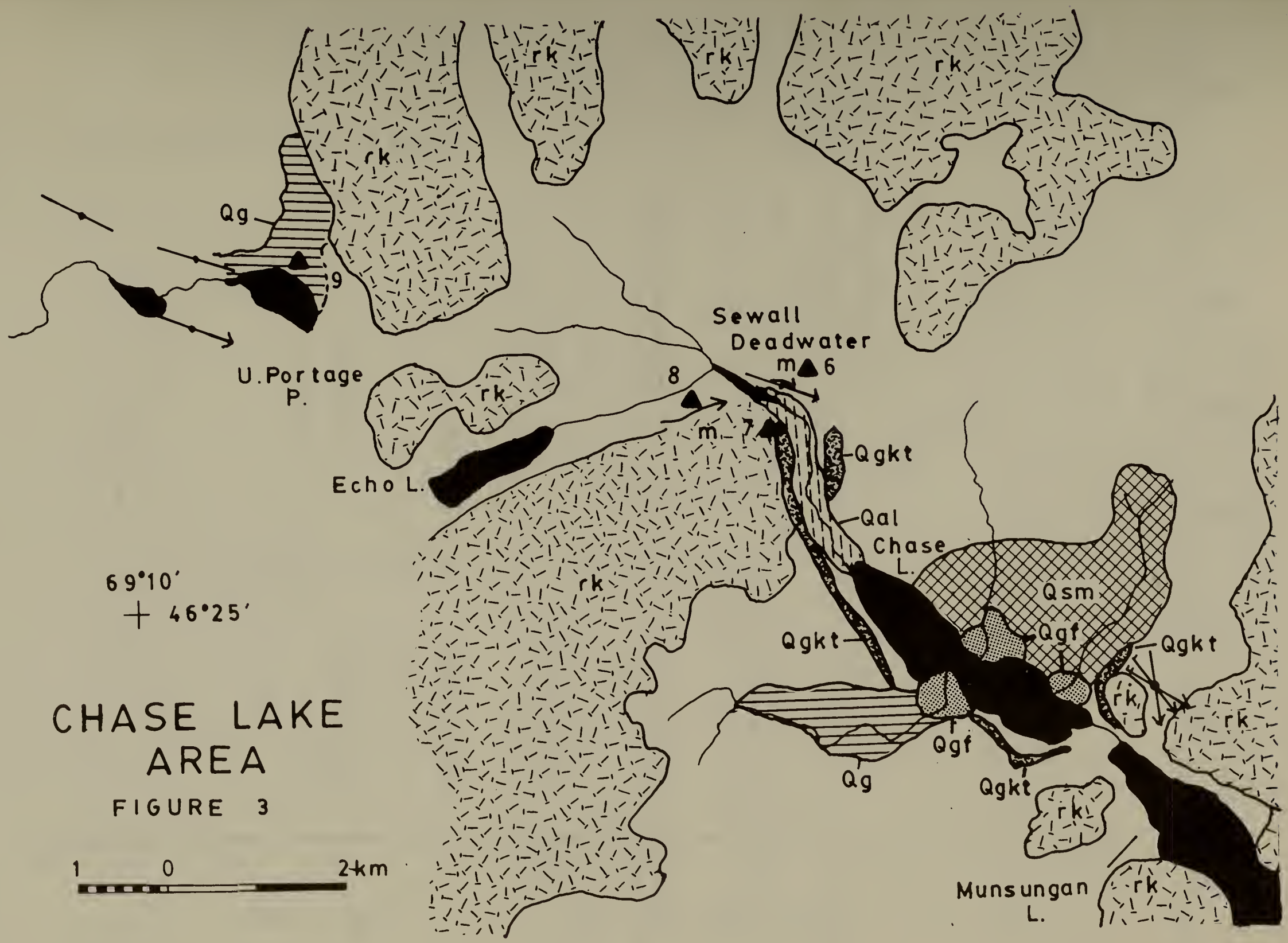
50.9 *Stop #8.* Meltwater channels.

This channel cuts into the south side of the Sewall Deadwater Pond gap at an elevation of 283 m. Several other channels also cut into this hillslope. From this location note the moderate relief to the north and west.

Interpretation: Stops 6,7, and 8. Hall (1970, p.42) noted the outwash surface and gravel from Upper Portage Pond (Stop 9) through Sewall Deadwater Pond, Chase Lake and into the Munsungan Lake Basin (Fig. 3). Further, he suggested the valley north of Echo Brook conducted meltwater from the Portage Ponds area into the Chase Lake basin. Due to the presence of outwash and delta gravels just above the present level of Munsungan Lake and the extreme depth of the lake (37 m), Hall suggested that stagnate ice blocks in the basin forced deposition along the ice margin.

Figure 3. Overleaf. Surficial map of the Chase Lake area. Symbols as Fig. 2. Additionally, Qal = alluvium, Qgkt = kame terrace, Qsm = stagnation moraine.

Figure 4. Second overleaf. Thalwag Profile of Upper Portage Pond to Chase Lake.



69°10'
46°25'

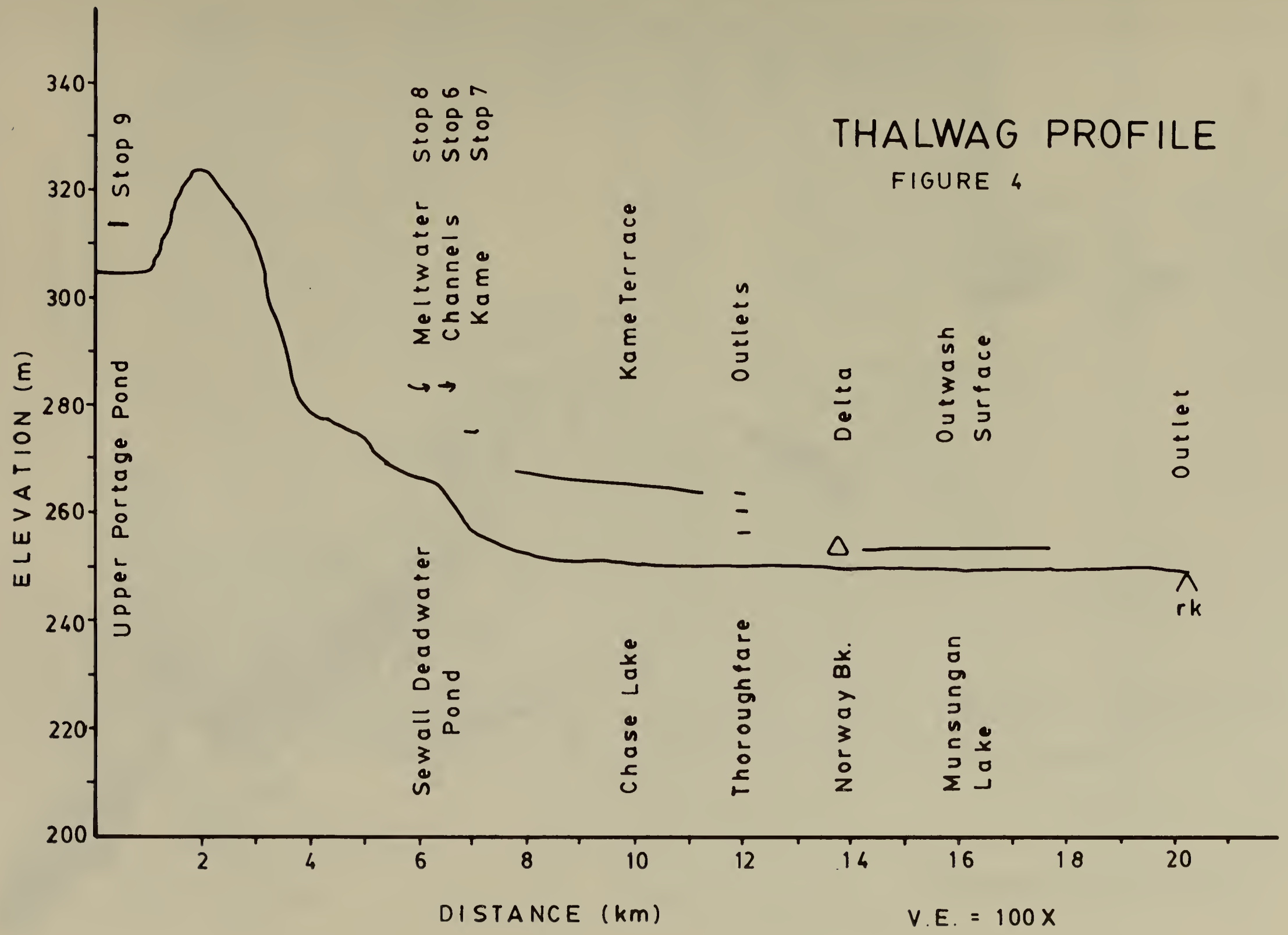
CHASE LAKE AREA

FIGURE 3

1 0 2 km

THALWAG PROFILE

FIGURE 4



A similar model applies to the Chase Lake basin except that the ice stagnated 15 m above the present water level. Several meltwater channels between Chase and Munsungan Lakes record the lowering of meltwater paths to the present level.

The hanging meltwater channels less than 20 meters above the Chase Lake terraces may be active as the kame deposit (Stop 7) and the terraces formed (Fig. 4). A lacustrine origin for these terraces seems unlikely in view of the relationships of the channels above Sewall Deadwater Pond and of the channels between Chase Lake and Munsungan Lake.

53.7 Keep right (W) on road to Churchill Lake at junction.

54.7 Keep right (N).

56.1 ***Stop. #9.*** Turn left (S) into gravel pit just before camp.

Some of the variable beds exposed here dip southward. This ice contact deposit has an elevation below that of the divide between Upper Portage Pond and Sewall Deadwater Pond.

Interpretation: The extensive gravels in this basin have not yet been studied. A preliminary interpretation is that the gravels above 330 m represent meltwater that was able to flow to the east through Sewall Deadwater Pond. The lower ice contact gravels may relate to opening of a drainage route west into the present Allagash Basin. Local topography plays the dominate role in controlling deglaciation.

Turn around, proceed back to junction of 53.7.

60.5 Proceed south on road toward Haymock Lake.

61.6 Pass road to left (E).

70.8 Rest stop at Haymock Campsite.

72.1 Turn right (W) and drive through gate.

75.7 ***Stop #10.*** Rat tail outcrop on west side of road.

Hard inclusions within the volcanic rock stood against the glacial erosion leaving the several rat tails see here. Ice flow was toward 103 degrees. The largest has more than 10 cm relief and a curious depression in the tail.

Optional Side Trip

The following directions are provided to access whaleback outcrops near the Allagash Waterway.

0.3 Cross Smith Brook (nice waterfall under bridge).

2.6 Cross Soper Brook.

3.8 Keep left past road to right (E).

5.3 Continue north through 4 way junction.

9.1 ***Stop #11.*** Whaleback outcrops exposed in old pit west of road.

Removal of the thin gravel revealed several whaleback forms shaped from the fine grained rock. These forms, up to 2 m wide by 10-12m long and 2 m high display fraction cracks and striations in the range of 110 degrees. Several local variations are also present.

9.6 John's Bridge over Round Pond.

Interpretation: Stops 10 and 11. Considering the quadrangle wide distribution of uniform ice flow indicators in the range from 100 to 115 degrees, it seems reasonable to assign this to the last erosive ice to cross this area. This region is somewhat unique in that this is the only area in northwestern Maine that shows a single consistent flow over a wide region. To the north indicators of this flow have northward indicators superposed on them. In the regions south and southwest, Hyland (1981) and I have noted that the east-southeast flows in the range from 150 to 180 degrees on top.

One interpretation places this area between the influence of downdraw in the St. Lawrence Seaway and the Gulf of Maine. Toward either ice stream ice flowed in response to the lowering profiles. However, between these lowering profiles the ice is essentially motionless and thus records no subsequent movement. The role of thermal regime controlling the distribution of these features is not yet known.

9.6 John's Bridge Allagash Waterway.

Turn around proceed south. On to Rip Dam !

79.3 Return to road near Haymock Lake turn right (S).

81.4 Continue south past junction with Pinkham Road.

82.0 BEWARE ! Moose crossing area.

82.5 Road right (W) to Indian Pond.

88.6 Town line T6 R11 (yellow post).

89.8 Road AW-7 right (N).

91.9 Cross Chamberlain Bridge. Note Mt. Katahdin to left (S).

92.0 Turn left (S) at parking lot.

95.1 Pass through North Maine Woods Telos Gate.

99.0 Enter Telos International Airport.

100.6 Telos Camp.

108.0 Pass west entrance Baxter State Park on left (E).

108.4 Keep left at Harrington Lake Dam Campsite.

115.2 Turn right (W) onto tar road after crossing the West Branch Penobscot River.

116.5 Turn right (NW) onto dirt road at Prays Camp sign.

116.9 Prays Camps, Ripogenus Dam; end of trip.

THE HURRICANE MOUNTAIN FORMATION MÉLANGE AND UNCONFORMABLY OVERLYING LOWER TO MIDDLE ORDOVICIAN VOLCANICS, BRASSUA LAKE AND MOOSEHEAD LAKE QUADRANGLES

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INTRODUCTION

Significance of mélange is central to development of accretionary tectonic theory despite the fact that dynamic processes internal to developing mélanges remain poorly understood. Inasmuch as the observation of mélange in the making is elusive, inference as to the origin of mélange in the geologic record — particularly the early Paleozoic — must be integrated from a diversity of structural and lithologic data. The purpose of this trip is to examine the Hurricane Mountain Formation within a logistically tractable, small portion of its regional extent. This necessarily restricts observation of the variety of clasts that occur in the unit throughout its strike belt, as well as of lithic variation in the matrix. Other varieties are described. Determination of depositional vs. tectonic modes of emplacement of clasts is hindered by subsequent structural modification during the Acadian orogeny, but nonetheless the early structural history of the Hurricane Mountain Formation must be gleaned in large measure from the modes of origin and emplacement of its clasts.

The relationship of pre-Silurian terranes bordering the Hurricane Mountain Formation remains speculative. Zen (1983) proposed that Taconian subduction took place along a lineament defined in part by the Lobster Mountain anticlinorium in which the Hurricane and its bordering pre-Middle Ordovician units lie. Geologic relationships in the Brassua Lake — Moosehead Lake part of the Lobster Mountain anticlinorium argue for inception of a mélange-forming event in later Cambrian, but not later than earliest Ordovician (Tremadocian) time. There is no evidence for further folding until the Acadian, inasmuch as Lower and Middle Ordovician (Boucot, 1969) and Upper Ordovician (Neuman, 1973) strata responded to Acadian deformation in virtually every respect similarly to Silurian and Lower Devonian strata overlying them. Together they form one structural package of singly-deformed and similarly-deformed rock. Similarity of pre-Middle Ordovician strata on each side of the Hurricane strike belt, projected northeasterly, as in the Caucomgomoc — Munsungun, and Weeksboro — Lunksoos anticlinoria, clouds the recognition of pre-Hurricane Mountain terranes, and heightens the probability that similar strata represent similar environments of deposition in different terranes, or that subsequent large-scale tectonic transport may have emplaced portions of one terrane upon another throughout

FIGURE 1 EXPLANATION

Silurian and Devonian

DSu — Silurian and Devonian metasedimentary rocks undivided

unconformity

Lower, Middle, and Upper Ordovician

Ok / Olm — Lobster Mountain Volcanics (Olm); mafic to intermediate volcanic and volcaniclastic rocks. Kennebec Formation (Ok); felsic volcanic and volcaniclastic rocks

unconformity

Upper Cambrian(?) to Lower Ordovician(?)

Cdr — Dead River Formation. Green slate, phyllite, and graywacke; commonly graded-bedded.

Cambrian(?)

Ch — Hurricane Mountain Formation. Gray to dark gray, rusty-weathering metasiltstone and subordinate phyllite, showing flaser bedding and complex (commonly hackly) microstructure. Cf. fig. 2 explanation for other lithic varieties.

GEOLOGIC OUTLINE MAP OF LOBSTER MOUNTAIN ANTICLINORIUM, EASTERN BRASSUA LAKE AND WESTERN MOOSEHEAD LAKE QUADRANGLES

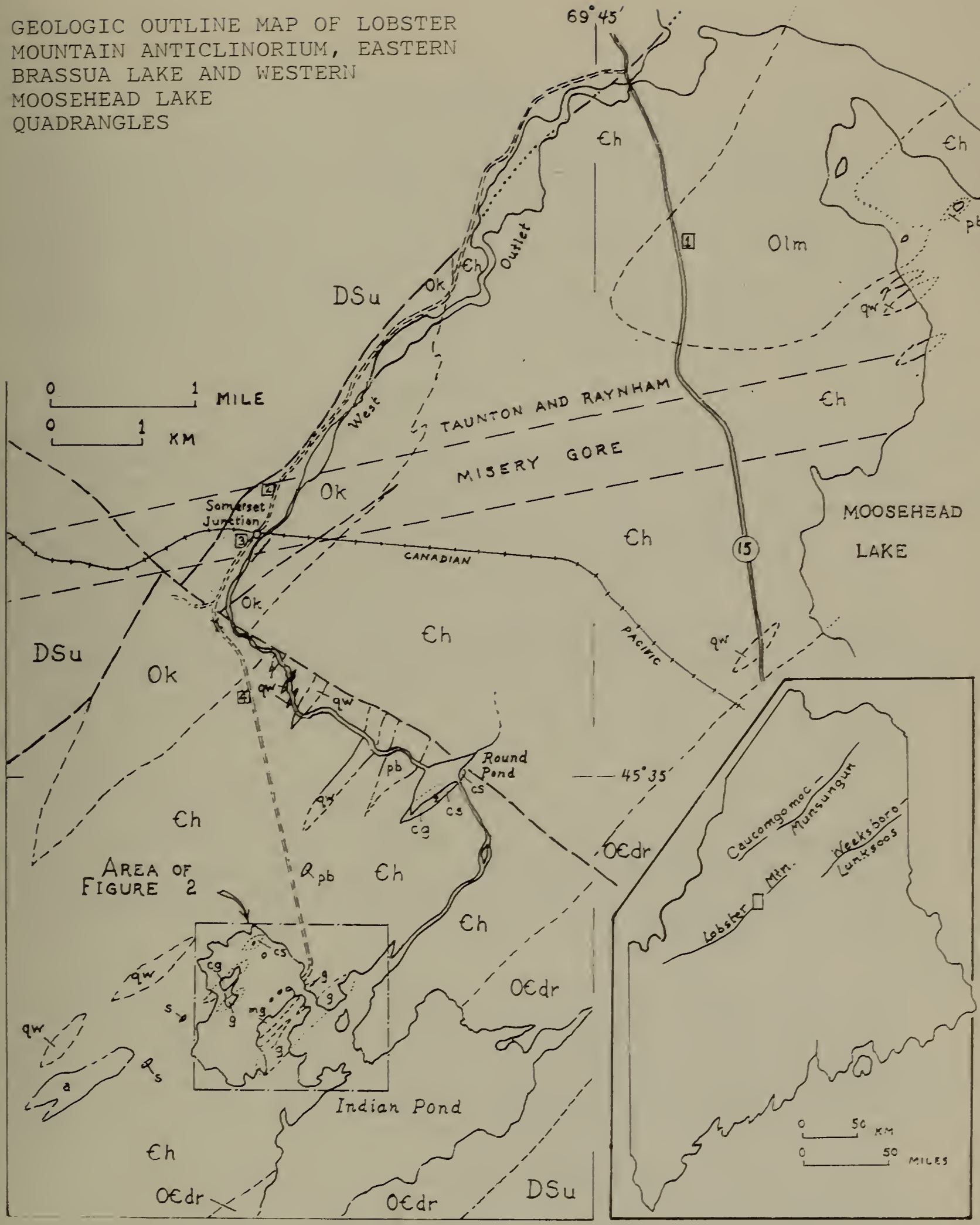


Figure 1

this part of the northern Appalachians.

ACKNOWLEDGMENTS

I thank the following who have patiently listened to, and contributed to my views on the Hurricane Mountain Formation and its significance: E.L. Boudette, B.A. Hall, C.E. Jahrling II, R.B. Neuman, S.G. Pollock, and P. St. Julien. I also acknowledge the field contributions of K.R. Frost in the Brassua Lake quadrangle in 1975, and those of a group of students of the 1980 Boston University field camp. K.R. Frost and Catherine Pflumio also contributed to laboratory study of this unit. Field support was provided by the Maine Geological Survey.

DESCRIPTION

The Hurricane Mountain Formation (Boone, in prep; Boone and Boudette, in prep.) forms a narrow outcrop belt extending from the vicinity of the Maine — New Hampshire boundary northeastward to the north end of the Lobster Mountain anticlinorium (fig. 1). It is lithologically distinctive from other pre-Silurian formations of the anticlinorium, and is interpreted as a tectonostatigraphic unit exhibiting characteristic features of mélange in many, but not all of its outcrop areas. Thickness of the formation, as measured in major limbs of Acadian folds, ranges from 850 to 1000 meters. This can be regarded only as a post-Acadian structural thickness.

Summary of contact relationships

The formation overlies the volcanigenic, less-deformed Jim Pond Formation of Cambrian age (Boudette, 1982; Boone, Boudette and Moench, 1981). It grades downward into thick olistostromal sequences of feldspathic quartzite and wacke of the Jim Pond in some areas, but in others, it apparently lies in fault contact with volcanics and volcaniclastic rocks assigned to the upper third of Jim Pond Formation.

The top of the Hurricane Mountain Formation is marked by gradation across thicknesses ranging from one or two meters to approximately 20 meters into the dominantly green, but color-veriegated slate and phyllite, and minor quartzite at the base of the Dead River Formation. Sparse graded beds of 10 - 30 cm thickness provide stratigraphic facing direction at the contact in several, widely dispersed localities in the Pierce Pond quadrangle (Boone, unpub. map; Lyttle, 1976) and in The Forks quadrangle (Burroughs, 1979). The contact is inferred to be diachronous younging to the southwest. The Dead River Formation is likely of late Cambrian to early Ordovician (pre-Arenigian) age; although pervasively polydeformed, it is not as intensely deformed as the Hurricane Mountain Formation. The Hurricane Mountain Formation is thus bounded by significantly less deformed pre-Middle Ordovician units.

Internal characteristics

The bulk of the formation consists of metasiltstone (about 70 percent), and slate or phyllite. This lithic mixture will be referred to as matrix, whether or not it is accompanied by clasts of internal or exotic origin. Both monomictic as well as polymictic assemblages are found in clast-bearing areas of outcrop. The question of whether these assemblages define structural or stratigraphic horizons within the formation is examined in a subsequent paragraph.

Matrix Conservatively estimated at least 95 percent of the matrix rock is devoid of well-defined bedding. Flaser-like pseudobeds appear as thin laminae and lenses ranging in length from a few centimeters to rarely more than a meter. These define a layering manifested by subtle differences in color and composition of metasiltstone and phyllite. Matrix may be gray, lacking sulfide and carbonaceous material, but commonly is dark gray, pyritic with consequent rusty weathering, and variably, but not intensely carbonaceous. Much smaller lenses, of a few millimeters to a centimeter in length, consist of light gray to buff, chalky weathering, finegrained felsic volcaniclastic rock. These are a common feature of the matrix, although such material perhaps accounts for a percent or less of matrix composition.

A structural characteristic of matrix not shared by clasts is a pervasive development of intersecting micro-fractures which endows the matrix with a hackly microstructure — one of the hallmarks of the formation. Where not healed by subsequent metamorphism or tectonism, the matrix is highly friable and is easily eroded.

Clasts Both endogenous and exotic rock types occur as clasts, rafts, and blocks within the mélange.

Endogenous clasts consist mainly of feldspathic sandstone and wacke. As quartz granules are characteristic of both types, the field term quartzwacke is used here in a collective sense. Variation in timing of lithification and disruption of clast materials relative to deformation of matrix resulted in a spectrum of geometric and structural relationships: from clasts of lithified quartzwacke with sharp, angular to rounded outline, to those that were probably semi-lithified at the time of emplacement (lumpy in outline, with gradationally sharp to diffuse boundaries), to quarter- to half-meter thick lenses showing pull-apart structure, to massive beds of quartzwacke. It is indeterminate whether the massively bedded areas themselves represent olistoliths or whether they represent detrital sediment deposited in situ.

Another possibly endogenous lithology consists of thinly bedded to laminated calcite-ankerite-rich metalimestone. The bedded carbonates show the same gradational relationships to matrix as does quartzwacke, but they are much less common. Rare fragments of cherty fine-grained quartzite also may be endogenous.

A rock type in the mélange that is difficult to class as being endogenous, exotic, or even as part of the Hurricane Mountain Formation, is massive brown slate with undeformed cleavage. Several outcrop areas of such slate, each commonly spanning a distance of 300 to 500 meters, have been found at widely separated localities along the strike belt of the formation. Each is surrounded by typical mélange matrix, but contacts have not been observed. Since matrix — as identified by its structural characteristics — rarely consists of slate unaccompanied by metasiltstone, these areas may represent rafts or blocks of slate in which the formation of cleavage predated disruption and emplacement. They thus may have been less prone to the same deformational response as unlithified matrix protolith sediments.[1]

Composition of exotic rock-types that have been found throughout the strike belt of the Hurricane Mountain Formation consist of:

volcaniclastic, arkosic sandstone

metagabbroic and metadiabasic rocks

pillowed and massive mafic metavolcanics

basaltic aquagene metatuff

pyritic, massive metacarbonate rock

amphibolite and greenschist

light gray thickly bedded orthoquartzite

black, carbonaceous orthoquartzite

serpentinite

felsic metavolcanics

chert

alkali granite

All but four have been recognized in the mélange in the Brassua Lake quadrangle; mafic varieties are commonest at Indian Pond. Here, as elsewhere, metagabbroic and metadiabasic rocks appear to occur both as intrusive bodies (dikes and sills) and as dismembered blocks. Evidence in support of each mode of occurrence will be examined.

[1] It is conceivable that the brown slate is a post-Lower Ordovician, pre-Acadian formation preserved only in widely separated synclinal keels, only within the Hurricane strike belt. Similar outcrop areas of green slate, however, are found in the Dead River Formation. It therefore seems more likely that each slate lithology is integral to its bounding formation despite the curious lack of evidence of polydeformation.

Pre-mélange metamorphism and deformation of metabasaltic rock

On islands of Indian Pond and surrounding mainland, blocks of tuffaceous greenstone, greenschist, and amphibolite show varied degrees of deformation that roughly correlate with their metamorphic grades. Insofar as is presently known, the ambient metamorphic grade of rocks in the Lobster Mountain anticlinorium, as recorded by white mica -- Fe-chlorite-rich assemblages in the Hurricane Mountain matrix and Dead River Formation, is lowest greenschist facies. No change from this assemblage, or change in grain-size of Hurricane matrix is detectable adjacent to the metabasaltic blocks showing higher grade metamorphic assemblages (Frost, 1977; Frost and Boone, 1978).

Fine-grained greenstone, which in some localities shows well preserved accretionary lapilli amid other primary textures of water-sorted tuff, is composed of the common assemblage

actinolite - chlorite - albite - quartz - Fe^{3+} oxide - pumpellyite or epidote $\pm$ white mica $\pm$ calcite

In one block at Round Pond (northeast of Indian Pond) assemblage (a) grades into rock containing the assemblage

actinolite - barroistic hornblende - chlorite - albite - epidote - quartz - hematite $\pm$ pumpellyite

Na in M4 is close to 0.75, from calculation of stoichiometry according to the method of Laird and Albee (1981) (Pflumio, 1983, unpub. ms.). Rocks containing this assemblage show a slightly coarser, porphyroblastic texture, with incipient tectonite fabric. According to Laird's (1980) grouping of oxide components, this is a c + 1 (and probably disequilibrium) assemblage (excluding pumpellyite). Similar chlorite-amphibole-bearing greenstone and greenschist occur on islands in the western part of the cove shown in figure 2. On one island, metabasalt dikes cut foliated greenstone and are themselves foliated. The foliation is approximately normal to that in the enclosing wall rock, and the mineral assemblage is dominated by relict pyroxene and plagioclase that is variably replaced by chlorite and unidentified (zeolitic ?) minerals, but in which actinolite and other amphiboles are lacking.

A block of metabasaltic rock 0.75 mile (1.3 km) long, 0.5 mile (0.8 km) west of Indian Pond is made up of amphibolite gneiss with highly contorted foliation and knots of pegmatitic albite. Alternating fine- and coarse-grained layers consist of the assemblage

Tschermakitic hornblende - albite - MgAl-rich chlorite - quartz - sphene - ilmenite.

Assemblages (a) through (c) record a transition between the biotite-albite and garnet-albite zones of a medium-pressure facies series (cf. Laird, 1980); a low-pressure series is excluded because a

progression from albite to oligoclase.....etc. is not recorded in the metamorphic progression to amphibolite gneiss.

Tectonic sequence vs. geographic restriction of clast types

The metabasaltic (greenstone to amphibolite) blocks of the Indian Pond area are roughly colinear in the middle of the Hurricane Mountain Formation belt, but other varieties of clast lithologies are not restricted to a tectonostratigraphic horizon. Quartzwacke clasts provide an example in the opposite extreme, for these are found at all positions relative to top and apparent base of the unit, and at all localities along its strike. Restriction to part of the strike belt may apply to some clast varieties as, for example, black and light gray quartzites in the southwestern Pierce Pond quadrangle, and most of the metabasaltic and metagabbroic varieties in the Brassua Lake quadrangle segment of the formation. Since the erosion surface provides only a two-dimensional examination of the unit, one may imagine that the total variety of clasts in a given geographic locale might be augmented were it possible to restore the section up-dip or examine the formation at depth. It is premature and probably false, therefore, to presume that the quartzites or the metabasaltic greenstones represent different provenances separated by the strike distance between each group of clasts.

Operating on the simple premise that clast variety along the total strike belt gives a reasonable integration, the observation may be made that striking similarities outweigh a few notable differences in the lithologies of the Hurricane Mountain Formation, the Chase Brook Formation (Hall, 1970), and parts of the Grand Pitch Formation (Neuman, 1962; 1967). Pollock (1982, and this volume) describes similar mélange characteristics in the Hurd Mountain Formation of the Caucomgomoc Lake area. The relationship between these units is presently the subject of further study.

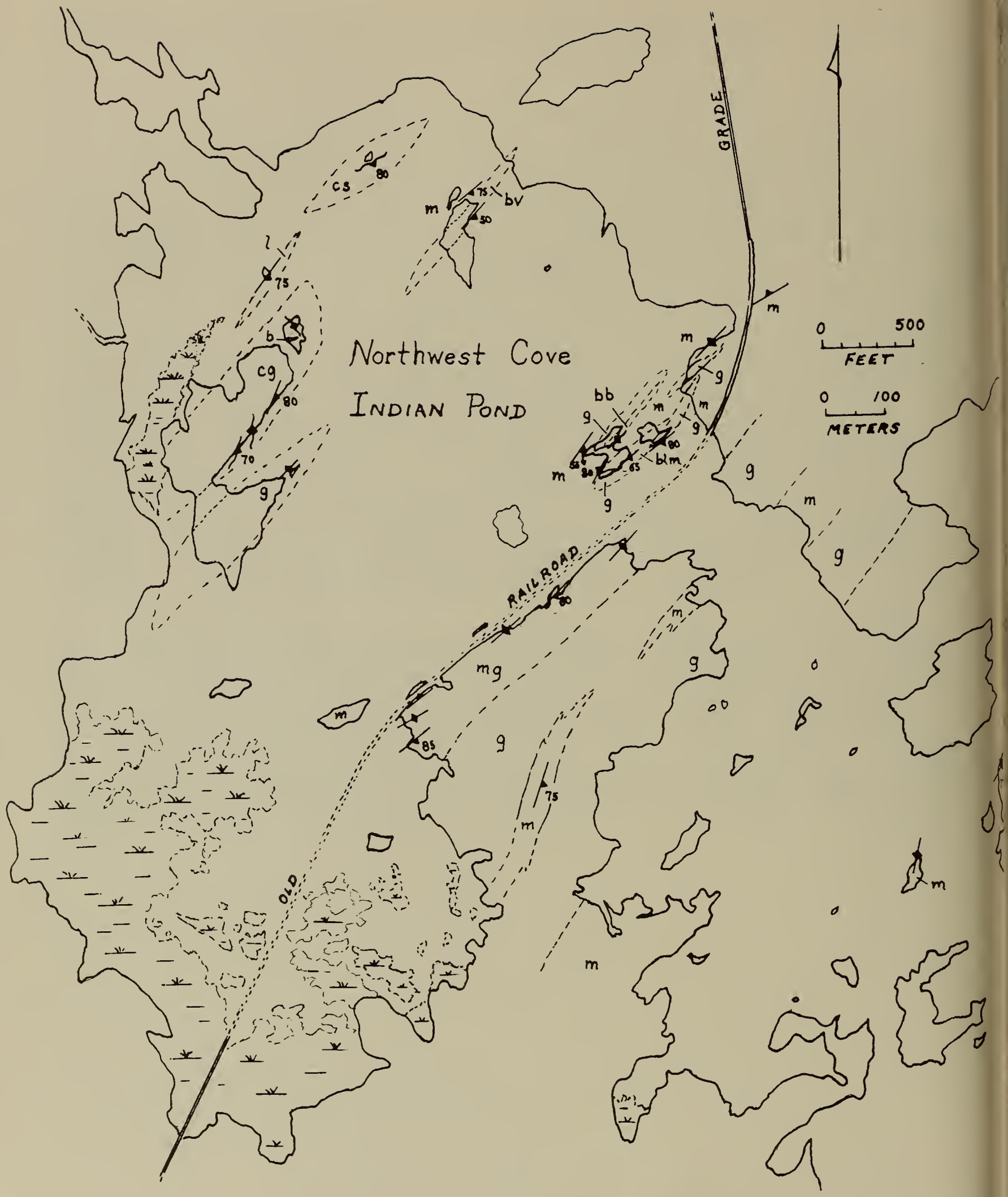


Figure 2. Lithologic relationships in Hurricane Mountain Formation, northwest cove of Indian Pond

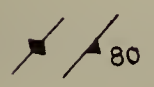
FIGURE 2 EXPLANATION

Hurricane Mountain Formation

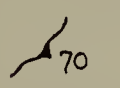
Cambrian (?)

m	Common matrix assemblage: rusty and non-rusty weathering, dark gray metasiltstone; black, light gray and light greenish gray phyllite
mg	Mélange matrix with abundant gabbro and basalt clasts, dikes, and sills
blm	Bluish gray metasiltstone
l	Rusty weathering, laminated chloritic metalimestone
b	Metabasalt
bv	Metabasalt, grading into bedded, calcitic, mafic volcaniclastics
bb	Metabasalt breccia
g	Fine - to medium - grained metagabbro
cg	Chlorite-actinolite greenstone and fine-grained greenschist; locally gradational into relict metabasalt
cs	Fine - to medium grained, well foliated, chlorite-actinolite-rich greenschist

Structure Symbols



Vertical or steeply dipping late cleavage in matrix; estimated average orientation of earlier, pre-mélange foliation in greenstone and greenschist





Contact; dashed where inferred



Bearing and plunge of small scale folds in matrix foliation

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Road Log, Trip

Moosehead Lake and Brassua Lake 15' quadrangles

Mileage Interval	Mileage Cumulative	
	00.0	Greenville Proceed west on State Rte. 15 toward Rockwood
13.0	13.0	Bridge crossing East Outlet of Moosehead Lake
3.8	16.8	Stop 1. Outcrops at crest of hill. Plagioclase-phyric volcanics assigned to Lobster Mountain and Kennebec Formation volcanic sequences
1.2	18.0	Bridge crossing West Outlet of Moosehead Lake
0.2	18.2	Turn left (south) on unpaved road. From here to Indian Pond, several access roads branch off westward; continue straight on main road, however, along the west side of the West Outlet
0.2	18.4 (approx.)	Enter Brassua Lake 15' quadrangle
3.8	22.2	Stop 2. Outcrops on west side of road, just south of Taunton & Raynham — Misery Gore twp. line. Lower to Middle Ordovician brachiopods recovered here from felsic metatuffs of Kennebec Formation (Boucot, 1969, p. 54-55)
0.5	22.7	Somerset Junction; Canadian Pacific R.R. overpass
0.05	22.75	Road intersection and bridge on left, but continue straight on main road
0.1	22.85	Stop 3. Typical exposures of cleaved felsic volcaniclastics of Lower to Middle Ordovician Kennebec Formation. Exposures

		on each side of Outlet show moderately dipping beds cut by steeply dipping spaced cleavage
0.25	23.1	Road intersection; keep left
0.2	23.3	Cross old railroad bridge
0.25	23.55	Well cleaved, dark red, green & white volcaniclastics of Kennebec Formation
		(From here to Indian Pond, road is in disrepair; proceed with caution)
0.3	23.85	Stop 4. Hurricane Mountain Formation. Typical matrix lithology, showing strongly cleaved, rusty weathering, dark gray metasiltstone and gray metaquartzwacke. Exposures of the contact between the Kennebec and Hurricane Mountain formations have not been found in the Brassua Lake and Moosehead Lake quadrangles, but 16 miles along strike northeastward, volcanics associated with the Lobster Mountain volcanic sequence overlie in exposed angular unconformity the Hurricane Mountain Formation. The Kennebec and Hurricane are inferred to be separated here by the same unconformity. Structural contrast between singly deformed Lower to Middle Ordovician rocks above, and polydeformed rocks of Cambrian to possibly Tremadocian age below the unconformity suggests the intervening early deformation is Penobscottian (cf. Neuman, 1967).
1.0	24.85 to 25.15	Abundant outcrops of Hurricane Mountain Formation; lithology similar to previous outcrops
1.0	26.15	End of road at east shore of northwest cove of Indian Pond. Weather permitting, embark by canoe to examine shoreline and island outcrops showing lithic variations and structural contrasts between mafic rocks (clasts and intrusives) and matrix of a mélange facies of the Hurricane Mountain Formation (map, fig. 2)
		Return to Greenville

GEOLOGY OF ROCKS ALONG THE EAST BRANCH PLEASANT RIVER, WHITE CAP RANGE, CENTRAL MAINE

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Introduction

The White Cap Range lies against the southern margin of the Katahdin Pluton and within a belt of contact- metamorphosed Upper Silurian and Lower Devonian metasedimentary rocks in the Kearsarge-Central Maine Synclinorium (Lyons, and others, 1982).

Three rock units are exposed along a 3 kilometer stretch of the East Branch Pleasant River from Trestle Brook to the base of Mud Gauntlet Falls (fig. 2). These units are tenatively correlated with the Madrid, Carrabassett, and Seboomook formations which lie along strike to the southwest.

Excellent exposures of the Carrabassett(?) and Seboomook(?) along the river exhibit a wide variety of sedimentary features which give insight into the mode of deposition, direction of sediment transport, and tectonic and depositional settings of these formations.

The Carrabassett(?) has an estimated thickness of 1800 meters(5900 feet) and consists of massive metapelite and medium- to thin-bedded metapelite with minor metasandstone. Nearly 80 percent (a rough estimate) of the rocks within the formation demonstrate signs of mass-gravity transport. Beds are folded, sheared, and/or broken into blocks. Similar, but less extensive, non-tectonic deformation in the Carrabassett has also been observed by the author 3.5 kilometers(2.5miles) south of Greenville on Route 15.

The contact between the Carrabassett(?) and Seboomook(?) is well exposed along Gauntlet Falls. Slumped beds of the Carrabassett are nearly vertical and form an undulating surface which is draped by the gently dipping beds of the Seboomook(?). This contact is discordant but may not represent a true unconformity.

The Seboomook(?) is composed of thin- to thick-bedded, poorly- to well-graded turbidites. Exposures within this section differ slightly from those of Seboomook at its type locality at Seboomook Dam (Boucot, 1961). The type section is characterized by well-graded, predominantly thin-bedded turbidites whereas at Gauntlet Falls the turbidites vary from thin- to thick- bedded over a relatively short stratigraphic distance. Nevertheless, Boone (personal communication) believes the Seboomook(?) at Gauntlet Falls is strikingly similar to rocks mapped by him on Little Poplar Mountain, east of the Carrabassett Valley, north of Kingfield (Boone, 1973).

Previous Work

Espenshade (1972) mapped the Moxie and Katahdin plutons and adjacent metasedimentary rocks between Greenville and Jo-Mary Mountain. Recent reconnaissance mapping by Mark Loiselle and Gary Boone in portions of the White Cap Range and adjacent areas was done in preparation for the new geologic map of Maine.

Acknowledgements

Dan Cochran and Lyman Feero, Woodlands Division, Great Northern Paper Company, kindly provided air photographs and maps of the field area. Mark Loiselle, Maine Geological Survey, introduced me to the geology of the region. Bill Betters and Dana Weinberg, Salem State, assisted in the field. I would like to extend special gratitude to Gary Boone who took the time to visit the section with me and offered invaluable advice and discussion.

Purpose of fieldtrip

The intent of this trip is as follows;

1. to introduce briefly the effects of the Acadian Orogeny along the western flank of the Kearsarge-Central Maine Synclinorium within and surrounding the White Cap Range;

2. to introduce the stratigraphy and structure of the area as demonstrated by a series of excellent exposures along the East Branch Pleasant River;

3. to study the sedimentary features and discuss modes of deposition, directions of sediment transport, tectonic and sedimentary environments of the Carrabassett(?) and Seboomook(?) in the area.

The majority of this trip will be devoted to studying and discussing the lithology and sedimentary features of the above formations which are exposed in several large outcrops along the East Branch Pleasant River.

Effects of the Acadian Orogeny

The rocks within this area were exposed to regional deformation and two metamorphic events during the Acadian Orogeny. Chlorite grade regional metamorphism and associated regional folding affected the region first, resulting in the development of upright to slighly overturned folds which strike and plunge northeast. In pelitic rocks folding was accompanied by the development of nearly vertical penetrative cleavage. This is best observed in rocks of the Carrabassett Formation which comprises a prominant slate belt extending from the White Cap Range westward to Greenville, and southwest through Monson.

In regions adjacent to the Katahdin and Moxie plutons contact metamorphism produced granofels and destroyed nearly all vestige of slatey cleavage. In the study area rocks have been prograded to hornblende-hornfels facies by the intrusion of the Katahdin pluton, exposed one kilometer (.6 mile) north of Stop 2. The extent of the contact aureole, greater than 5 kilometers, implies that the closest contact with the pluton may not be to the north but at depth. The contact metamorphism of this area is strikingly similar to that of the Skiddaw Aureole, Cumbria, England (Rastall, 1910).

Pelitic beds and laminae have developed medium- to coarse-grained granoblastic and porphyroblastic textures in which cordierite, andalusite and biotite can often be recognized in hand specimen. Feldspathic and quartz-rich psammitic beds have undergone less change, are generally finer grained, and lighter in color. The resultant contrast between the metapelitic and metapsammitic beds and laminae enhances the visability of laminations and grading of beds, which in regions of pervasive cleavage, are difficult to distinguish.

Stratified Rocks

Three units are exposed along the East Branch Pleasant River. These rocks can be tenatively correlated with the Madrid(Moench, 1971), Carrabassett (Boone, 1973), and Seboomook(Boucot, 1961 and Boone, 1973) formations along strike to the southwest on the basis of resemblence and stratigraphic relationships. However, lacking paleontologic evidence, correlation at this point is speculative.

Madrid(?) Formation

The Madrid(?) along the East Branch Pleasant River is exposed in the core of an anticline and along the Mud Gauntlet Fault (here named) where it is brought up against the Seboomook(?). Only the upper portion of the formation is exposed in both locations.

Where exposed, in this area, the formation is a gray to purple-gray biotite metasandstone. The unit is predominantly massive and bedding is at best faint. Bedding is observed near the contact with the overlying Carrabassett(?) where faint, slightly contorted laminations are visable. One of the distinguishing characteristics of this unit are the numerous greenish-gray calcsilicate pods scattered throughout it.

Carrabassett(?) Formation

The Carrabassett(?) has an estimated thickness of 1800 meters (5900 feet) and consist of alternating units of massive metapelite with thin- to medium-bedded, generally poorly-graded metapelite and metasandstone, with rare lenses and blocks (olistoliths?) of light-gray, quartz-rich metasandstone one to two meters thick. The sandy portions of graded beds are generally less than 20 percent but locally may be as high as 40 to 50 percent. On the fresh surface the metapelite appears dark purple and glittery. Porphyroblasts of andalusite and cordierite are commonly visable.

The weathered surface is characteristically pitted and rough because of the removal of cordierite and the resistance of andalusite, which stands out as small chalky-white knots or layers.

Mass-gravity transport appears to have played a major role in the deposition of the formation in this area. Flow(?) and incoherent slumping affected nearly 80 percent of this unit. At the moment is is difficult to state conclusively whether some of the massive metapelites were indeed massive, or simply well-bedded, predominantly pelitic units which became homogenized by subaqueous flow.

Mass-gravity transport within the Carrabassett(?) of this section is evidenced by;

1. folding and shearing of bedding which does not conform to the the local structure;

2. undisturbed beds which are sandwiched between deformed beds of the same composition;

3. angular to subrounded blocks of bedded metapelite, metasandstone and massive metapelite are incorporated in a deformed matrix of the same or slightly different composition.

The following features, preserved in the metasediments provide insight into the source and directions of sediment transport during deposition.

1. **Slump Folds**

 a. Converging direction of folds.

 Most of the slump folds appear chaotic (fig. 4) and no sense of motion has been conclusively determined. At Stop 3B the fold hinges are exposed and give the appearance of a southward directed slump.

 b. Orientation of fold hinges

 Hinges of the slump folds (Stop 3B) are roughly parallel to the local tectonic trend. This may be coincidence or suggest mass-movement into a tectonically controlled basin or trough.

2. **Minor faulting of beds**

 Small gravity faults suggest slumping of sediment toward the south (Stop 4).

3. **Current structures**

 Apparent current directions, determined from ripple foresets, indicate transport toward the ENE, parallel to the local tectonic trend.

4. **Dip of the formation**

Undisturbed beds within the formation have southerly dips 20°-30° steeper than those of the overlying Seboomook(?). This may suggest tilting prior to deposition of the Seboomook and/or a steeper south-facing depositional slope.

The lower contact of the Carrabassett(?) and the underlying Madrid(?) is not exposed but can be inferred to within a few meters. The fine laminae in the Madrid(?) are contorted near the contact which is overlain by a massive metapelite unit of the Carrabasset(?).

The upper contact is well exposed at Gauntlet Falls. The slumped beds of Carrabassett(?) are highly contorted and nearly vertical. The uneven surface formed by the Carrabassett(?) is draped by a well-bedded, undisturbed, homoclinal sequence of Seboomook(?) metasediments.

Although the upper and lower contacts are discordant they do not necessarily imply an interruption in sedimentation and may not be true unconformities.

The Carrabassett(?) along the East Branch Pleasant River differs slightly from that at its type locality in the Carrabassett Valley(Boone, 1973). The quartzwacke member does not occur in this section, however the quartzwacke occurs as discontinuous lenses and is commonly missing from local section. In the type locality the massive metapelite underlies a thinly-bedded member. In this section the massive metapelite and thinly-bedded units alternate and appear to have no fixed stratigraphic position. The upper, discontinuous, variegated, rusty-weathering calcsilicate member (Hildreth Formation of Osberg and others, 1968) is missing along the East Branch Pleasant River. However, a unit strikingly similar to this member is exposed 8 kilometers(5 miles) to the northwest, along the south end of Upper Jo-Mary Lake (see mile 20.3 of road log).

Seboomook(?) Formation

Between Gauntlet and Mud Gauntlet Falls there is 550 meters (1800 feet) of nearly continuous outcrop. A thickness of 300 meters(985 feet) of Seboomook(?) is exposed along this section and can be studied in detail from the basal contact at Gauntlet Falls, to Mud Gauntlet Falls where the section is truncated by a fault along which Madrid(?) is brought up(fig. 3).

The base of the Seboomook (Fig. 5) is characterized by a sequence of thin-bedded turbidites consisting of thin ripple trains of metasandstone overlain by thick metapelite (CE divisions, Bouma, 1963). There is no visable grading from the metasandstone to the metapelite or within the metapelite. The thin-bedded unit changes rapidly to thick-bedded turbidites (AB divisions, Bouma, 1963) and back to thin-bedded over a relatively short stratigraphic thickness.

This rapid change in sequences has been seen at other exposures in the area and may represent levee or basin-plain deposits overlain by those of a prograding suprafan lobe (Walker, 1976).

Superficially these rocks resemble the Seboomook at its type locality northwest across strike at Seboomook Dam(Boucot, 1961). However, these rocks lack the well-graded, monotonous, cyclic bedding characteristic of the type section. Nevertheless, the Seboomook(?) along the East Branch Pleasant River closely resembles turbidites, mapped as Seboomook(Boone, 1973) on Little Poplar Mt., along strike to the southwest. Sediment transport within the Seboomook(?) in this area is principally from the south (figs. 5 and 6) and not from the east as noted in the Seboomook to the northwest (Hall and others, 1972, 1976).

Local Structure

The section of the East Branch Pleasant River which will be visited extends from the core of an anticline, exposing Madrid(?), 2.4 kilometers upstream from Gauntlet Falls, to the axis of a faulted syncline at Mud Gauntlet Falls. The Madrid(?) and slices of the Carrabassett(?) have been juxtaposed against the Seboomook(?) along the fault. The total thickness covered will be approximately 2250 meters(7380 feet). This figure also represents the approximate amount of stratigraphic displacement which has occurred along the Mud Gauntlet Fault. This fault connects with a major thrust which extends southwest through Wyman Lake, near Bingham, southwest toward the Maine-New Hampshire border (Boone, personal communication). To the Northeast the fault extends for 16 kilometers (10 miles) where it is truncated by the Katahdin Pluton.

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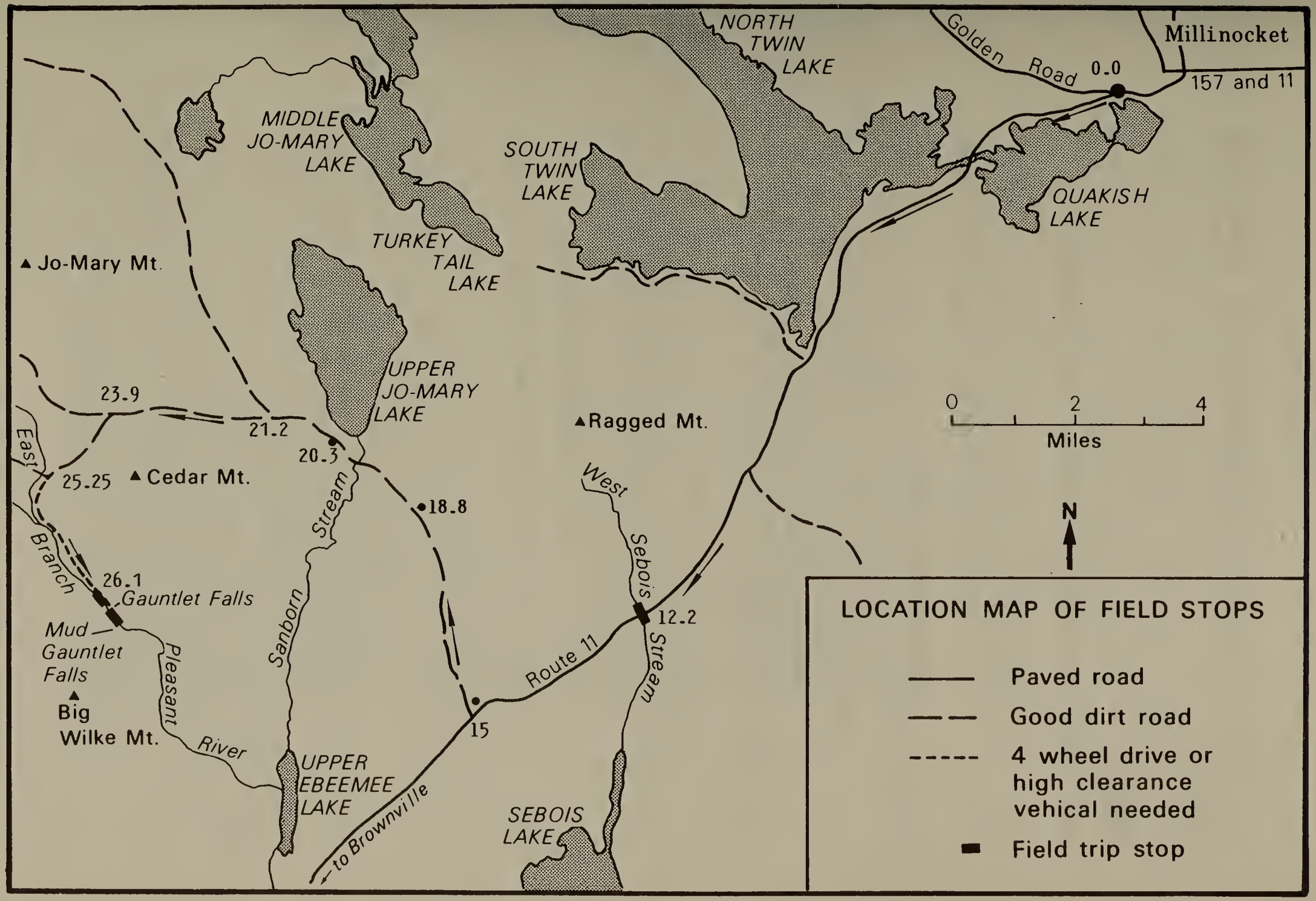


Figure I. Location map of areas visited. Important road-log mileages are shown above. Road log begins at the junction of the Golden Road and Route II (mileage 0.0 on map).

Road Log

This road log begins at the Junction of Rt. II (Brownville Road) and the Golden Road (Great Northern Road). For a map showing areas visited refer to figure I.

Mileage

0.0 Head south on Route II toward Brownville.

3.0 Cross over the West Branch Penobscot on green bridge.

6.7 Cross railroad tracks near Partridge Cove Marina.

12.2 **Stop I: Sebois Stream**

An excellent exposure of the thinly-bedded member of the Carrabassett(?) lies in the stream beneath and on both sides of the bridge. These rocks have not been affected by contact metamorphism and exhibit the low-grade regional metamorphism and pervasive cleavage which characterize Acadian deformation in this region.

The bridge stradles the axis of a slightly overturned, northeast-plunging syncline. Beds on the north side strike N50°E and dip 84°S with tops toward the south. Beds on the south side of the bridge strike N55°E and dip 89°S with tops toward the north. Bedding and cleavage are parallel. The nose of the syncline is visable beneath the bridge.

15.00 Turn off Rt. II onto dirt road heading for Jo-Mary Lake Campground.

Spotted hornfels

On the northwest corner of this intersection is a small outcrop of medium-gray spotted hornfels. The outcrop lies on the outer margin to two contact areoles; that of the Schoodic Pluton to the south and the Katahdin Pluton to the north. The rocks no longer exhibit slately cleavage but still maintain signs of foliation.

18.80 **View of Jo-Mary Mountain**

Jo-Mary Mountain comes into view toward the northwest. The mountain is composed of the same granofels we will be studying along the East Branch Pleasant River. The valleys surrounding the mountain are underlain by portions of the Katahdin and Moxie plutons.

19.8 Road enters and leaves gravel pit.

20.30 **Upper member of the Carrabassett(?) Formation**

A rusty-weathering, banded, calcsilicate is exposed on the left-hand (southwest) side of the road. These rocks may be the upper member of the Carrabassett(?) (Boone, 1973, also known as the Hildreth Formation of Osberg and others, 1968). Compared to the surrounding rocks, these are highly folded. The upper member is generally thin and commonly responds to deformation by disharmonic folding.

20.66 The road to Jo-Mary Campground joins from the right(northeast). Do not turn, continue straight.

21.20 Road splits. Take lefthand fork. Continue straight, passing three gravel-based logging roads which enter from the left(south).

23.90 Road splits. Take lefthand fork toward Jo-Mary Pond.

24.40 Pass Jo-Mary Pond (not shown on map).

24.70 Road from Cedar Mountain joins from the south. Continue straight.

25.25 Turn left(south) on bumpy-crunky road to Gauntlet Falls. People having vehicles with low clearance must park here and hitch a ride.

26.10 **Stop 2: Madrid(?) Formation**

Park in clearing on west side of road.

The remainder of the trip will be devoted to looking at the rocks along the East Branch Pleasant River. A detailed map illustrating the geology and stops along the rivers is presented in figure 2.

An exposure of Madrid(?) lies along the east bank of the river west of the road. The rock is a purple-gray, biotite metasandstone with greenish-gray calcsilicate pods. The Madrid(?) and the massive metapelite at the next stop appear very similar on the fresh surface. However, the metasandstone lacks the cordierite pits and resistant porphyroblasts of andalusite which characterize the weathered surface of the metapelite.. Very faint laminations can be seen on the south end of the outcrop near the bank.

26.40 **Stop 3: Contact between Madrid(?) and Carrabasset(?)**
Slump folds

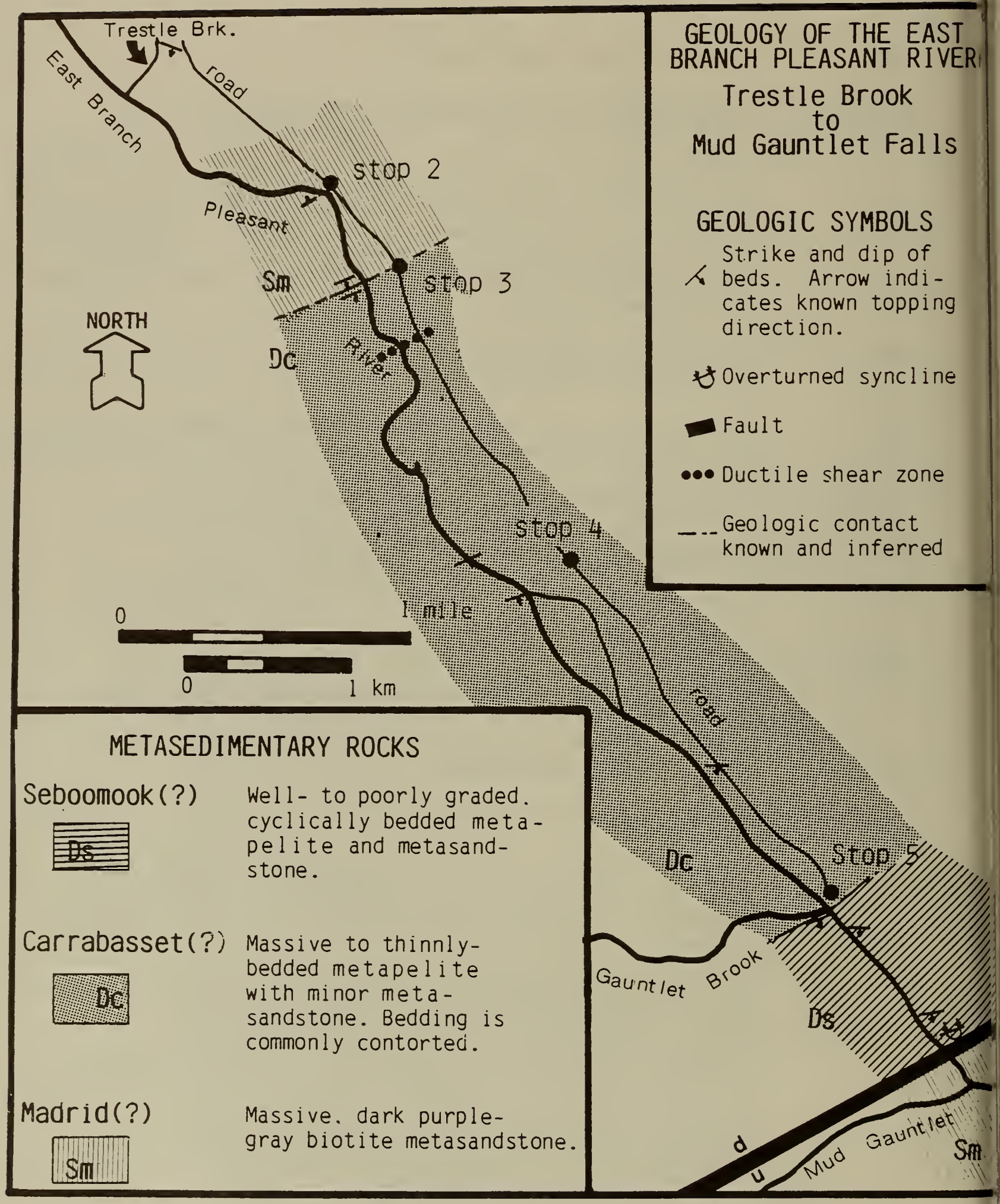


Figure 2. Geology and location of stops along the East Branch Pleasant River.

Park where possible allowing enough room for other cars to pass. Walk due west to the river, approximately 60 meters.

Stop 3A: Contact

Extending from the northwest shore are several small strike ridges of Madrid(?). Bedding is faint and often contorted. The contact lies south of these ridges and is covered with gravel.

The next set of exposures, south of the contact, are massive and bedded metapelites of the Carrabasset(?) Formation. The massive metapelite is best exposed on the east bank. The weathered surface of the granofels is highly pitted and andalusite crystals up to 1 cm long are visable. Overlying the massive metapelite is a bedded pelitic unit (best exposed on the west bank) containing thin beds of starved ripples. The apparent paleo-current indicated by ripple foresets is toward the ENE, parallel to the local tectonic trend.

Stop 3B: Slump Folds

Walk downstream to the next set of large outcrops(approximately 100 meters) and cross to the west bank. Looking across to the outcrop on the east bank several large slump folds are visable. The axes of the folds are exposed giving the outcrop a rounded appearance. The fold hinges appear to thicken toward the south and strike E-W.

Return to the road following the same route back. **Do not** head due east toward the road or you **will** be lost in an alder swamp.

26.90 **Ductile Shear Zone**

An excellent exposure of a ductile shear zone lies about 30 meters to the west of the road, along the east bank of the river. The shear zone cuts a bedded unit of the Carrabassett(?). Differentiation has occurred along closely spaced shear planes oriented at a slight angle to bedding.

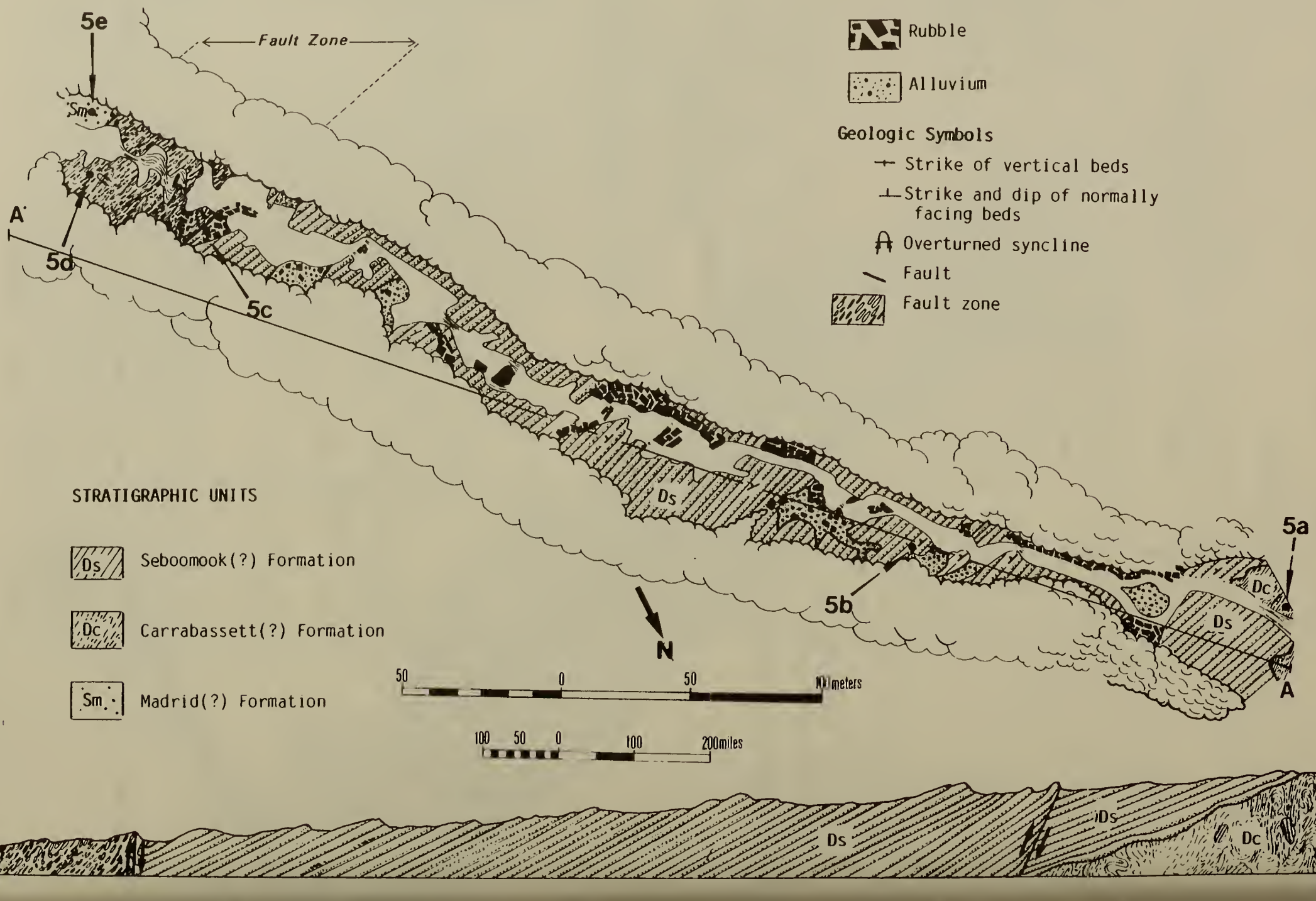
27.30 **Stop 4: Metasandstone lenses and/or Olistoliths(?)**

Park where possible. Walk 300 meters due west of the road to the river.

A well-bedded unit of the Carrabassett(?) is exposed in several large outcrops along the river. Many of the beds are 30-50% psamitic. Beds are folded and cut locally by numerous small gravity faults. A discontinuous bed and numerous transported blocks of massive, light-colored, quartz-rich meatsandstone are present. These beds are not characteristic of the Carrabassett(?) and may have been transported from a region of more proximal deposition.

Geology of the East Branch Pleasant River along Gauntlet and Mud Gauntlet Falls, TBRIO

(Sketch map controlled by pace and compass traverse)



Vegetated ledge

Rubble

Alluvium

Geologic Symbols

- Strike of vertical beds
- Strike and dip of normally facing beds
- Overturned syncline
- Fault
- Fault zone

STRATIGRAPHIC UNITS

- Ds — Seboomook(?) Formation
- Dc — Carrabassett(?) Formation
- Sm — Madrid(?) Formation

27.80 **Stop 5: Contact between Carrabassett(?) and Seboomook(?)**
Detailed look at Seboomook(?) Formation
Overturned syncline and Mud Gauntlet Fault

Park in the large clearing at the head of Gauntlet Falls. For this stop consult the sketch map of the river from Gauntlet to Mud Gauntlet Falls (fig. 3).

Along both sides of Gauntlet Falls is a spectacular series of exposures showing the contact between the Carrabasset(?) and Seboomook(?) formations.

Stop 5A: Rocks near the Carrabasset(?) - Seboomook(?) contact

Walk across the river at the head of the falls.

The Carrabassett(?) is best exposed on the west side. On the water-worn surface the rock is light- to medium-gray, and bedding can be easily seen where present. The unit is dark- to medium gray and appears massive on highly weathered surfaces and where blasted. Bedding is highly contorted where visable. Blocks ranging from a few centimeters to more than a meter in length lie in a surrounding matrix of metapelite. Some blocks are composed of massive metapelite or sets of thin, well- to poorly-graded beds.

Looking across to to the east wall of the bedrock channel a faint, complex fold pattern is visable(figure 4). This exposure illustrates nicely the type of non-tectonic deformation which is so characteristic of the Carrabassett in the area.



Figure 4. Slump folds in the Carrabassett(?) Formation at Gauntlet Falls.

The Seboomook(?) above the contact consist of thin-bedded, poorly-graded turbidites(fig. 5). The Metapsamitic beds are less than 1cm thick and consist of starved ripples. The intervening metapelites are up to 15cm thick and show no grading. These beds of Seboomook would be categorized as CE divisions of the Bouma sequence and may represent back levee, levee, or basin-plain deposits.

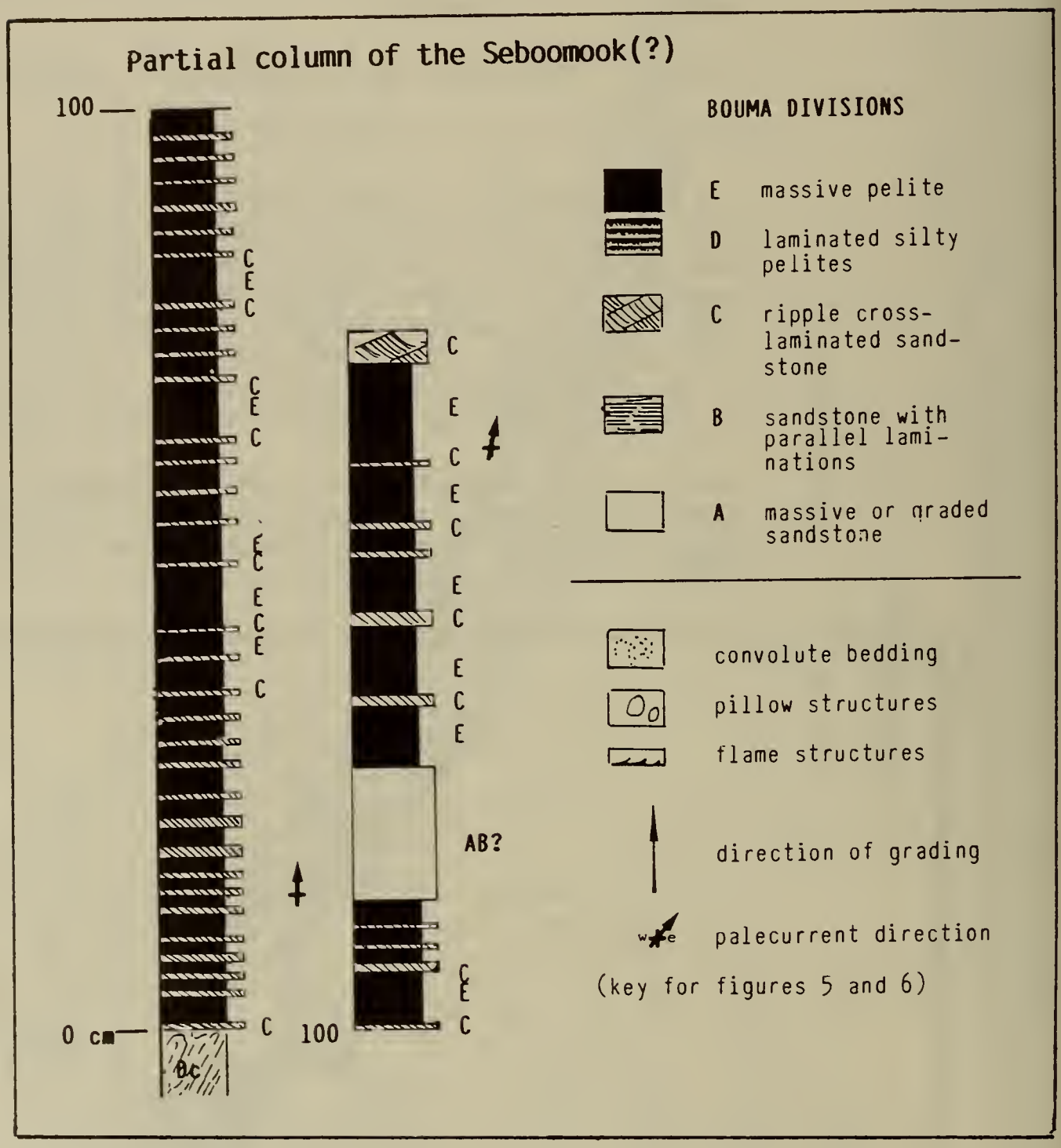


Figure 5. Thin-bedded turbidites overlying the Carrabassett(?)-Seboomook contact. Most of the original sedimentary structures have been preserved even though these rocks are now granofels. Preservation of these structures makes it possible to describe the turbidites according the the Bouma Divisions(Bouma, 1962) given above. The turbidites at the base of the Seboomook consist of thin, starved-ripple trains interlayered with massive pelite (CE).

Stop 5B: Thick-bedded unit of the Seboomook(?)

The thin-bedded turbidites which lie over the contact change quickly to thick-bedded, dominated by AB and BC sequences sequences (fig. 6). These rocks may represent a prograding suprafan lobe. The Seboomook changes back to thin-bedded before it is truncated by the fault.

Paleocurrents as determined from ripples are toward the north.

Stop 5c: Overturned syncline and beginning of fault zone

A fault on the south limb of an overturned syncline is exposed along a strike ridge approximately 5 m wide. This fault is the northeastern-most fault in the Mud Gauntlet Fault zone. The fault cuts the southwestern limb of an overturned syncline and brings up Carrabassett(?)-like rocks against the Seboomook(?). Beds on the northeast side of the fault are thin, well- to poorly graded, and strike N80°E. These beds are overturned, dipping 80° SE with tops toward the NW.

Stop 5D: Overturned syncline in fault zone

Within the fault zone is another overturned syncline, exposed on a water worn pavement. The fold is recognized by a thick, 15cm, metasandstone bed. Graded beds on the south side have tops toward the NE and those on the north side have tops to the South.

Stop 5E: Madrid on south side of fault

South of the fault zone lies the light to dark, purple-gray, biotite metasandstone observed at Stop 2. The total stratigraphic displacement along the fault is estimated at 2250 meters (7380 feet).

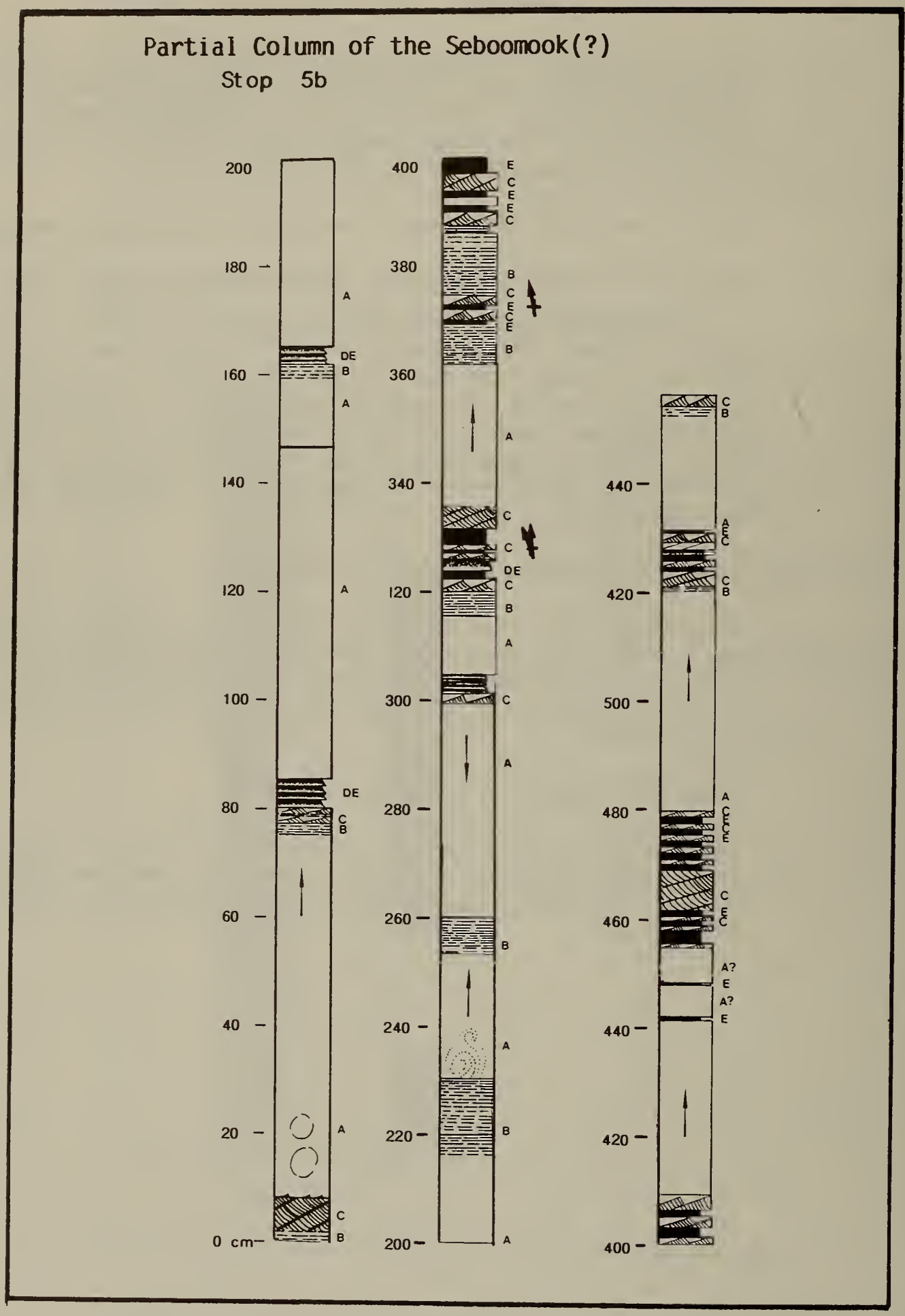


Figure 6. Thick-bedded turbidites of the Seboomook(?) exposed at stop 5b. In this section A and B divisions are much more prevalent and may indicate a prograding suprafan(Walker, 1976).

TRIP B-3

FEATURES OF UPPER SILURIAN AND LOWER DEVONIAN SEDIMENTARY ROCKS IN THE CAUCOMGOMOC LAKE AREA NORTHWESTERN MAINE

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INTRODUCTION

The purpose of this field trip is to examine the sedimentary structures and lithologic variability and to interpret depositional environments in two formations in the Caucomgomoc Lake area. The first of these has not been formally named or described and will be referred to informally as the Allagash Lake Formation[1]. This formation is of Late Silurian (Pridoli) and Lower Devonian (Gedinnian) age. The second formation is the Seboomook Formation of Lower Devonian (Siegenian) age. The sedimentary rocks of these two formations are distinctively different in terms of lithologies, sedimentary structures, and interpreted depositional environments.

AGE AND CORRELATION

Prior to work by Pollock (1983), no systematic mapping effort had been made in the Caucomgomoc Lake and Allagash Lake quadrangles. Reconnaissance in the 19th century was by Jackson (1838), and Hitchcock (1861). Hitchcock (1901), considered the volcanic rocks of the Pre-Silurian (figure 1) and Allagash Lake Formations to be "...'trappean and altered' rocks on Lakes Allequash (and) Cauquomogomoc...". He considered the sedimentary rocks of both the Allagash Lake and Seboomook Formations to the "Silurian and Cambrian Clay Slate." Keith (1933), followed essentially the same usage as established by Hitchcock. Doyle (1967), in his compilation of the area essentially showed correct distribution of the sedimentary rocks assigning a Lower Silurian (Upper Llandovey) to Lower Devonian (Siegenian) age for sedimentary rocks of the Allagash Lake Formation and an Ordovician to Silurian age for the volcanics. A Lower Devonian (Upper Gedinnian and Siegenian) was assigned for sedimentary rocks of the Seboomook Formation. Sprague (1972), partially mapped the Caucomgomoc Lake

[1]The name Allagash Lake Formation has been approved for use by the Geologic Names Committee of the U.S. Geological Survey. No attempt is made to formally describe this unit here. As used here, this name is strictly informal and its use is only intended to facilitate discussion. Formal naming and description will be done elsewhere. Informal usage is with permission of the Geologic Names Committee.

EXPLANATION

LOWER DEVONIAN

Dsu — SEBOOMOOK FORMATION UNDIFFERENTIATED

ORDOVICIAN(?) TO DEVONIAN(?)

ODf — FRONTENAC FORMATION

UPPER SILURIAN AND LOWER DEVONIAN

ALLAGASH LAKE FORMATION

DSav - predominantly pillowed baslat with interbedded conglomerate, limestone and sandstones of various classification

DSas - mixed sedimentary rocks including limestone, wackes, arenites, siltstones and shales (with and without hematite cement) and conglomerates.

CAMBRIAN(?) AND ORDOVICIAN(?)

UNIT 3 (Hurd Mountain Formation)

€Ohb - pillow basalts, basaltic agglomerate and basaltic lapilli tuffs

€Ohm - meta-siltstone, claystone slate and phyllite minor calcareous meta-siltstone. Commonly

€Ohs exhibits rusty weathering. Unit is complexly deformed and pervasively sheared. Unit is interpreted as a tectonic melange. Other rock types in the pelitic host include quartz wacke, meta-basalt, ultra-mafics, gabbros, diorites and granodiorites.

€Ohs - medium to thick bedded quartzose wacke, usually texturally uniform and lacking sedimentary structures.

UNIT 2 (Caucomgomoc Lake Formation)

€Ocb - a dominantly phyric and aphyric meta-basalt flows undifferentiated.

$€Ocb_2$ - upper member on Caucomgomoc Lake. Phyric and aphyric pillowed and non-pillowed basalt flows, basaltic agglomerate and basaltic lapilli tuff.

€Ocs - medium to thick bedded quartzose wacke and siltstone.

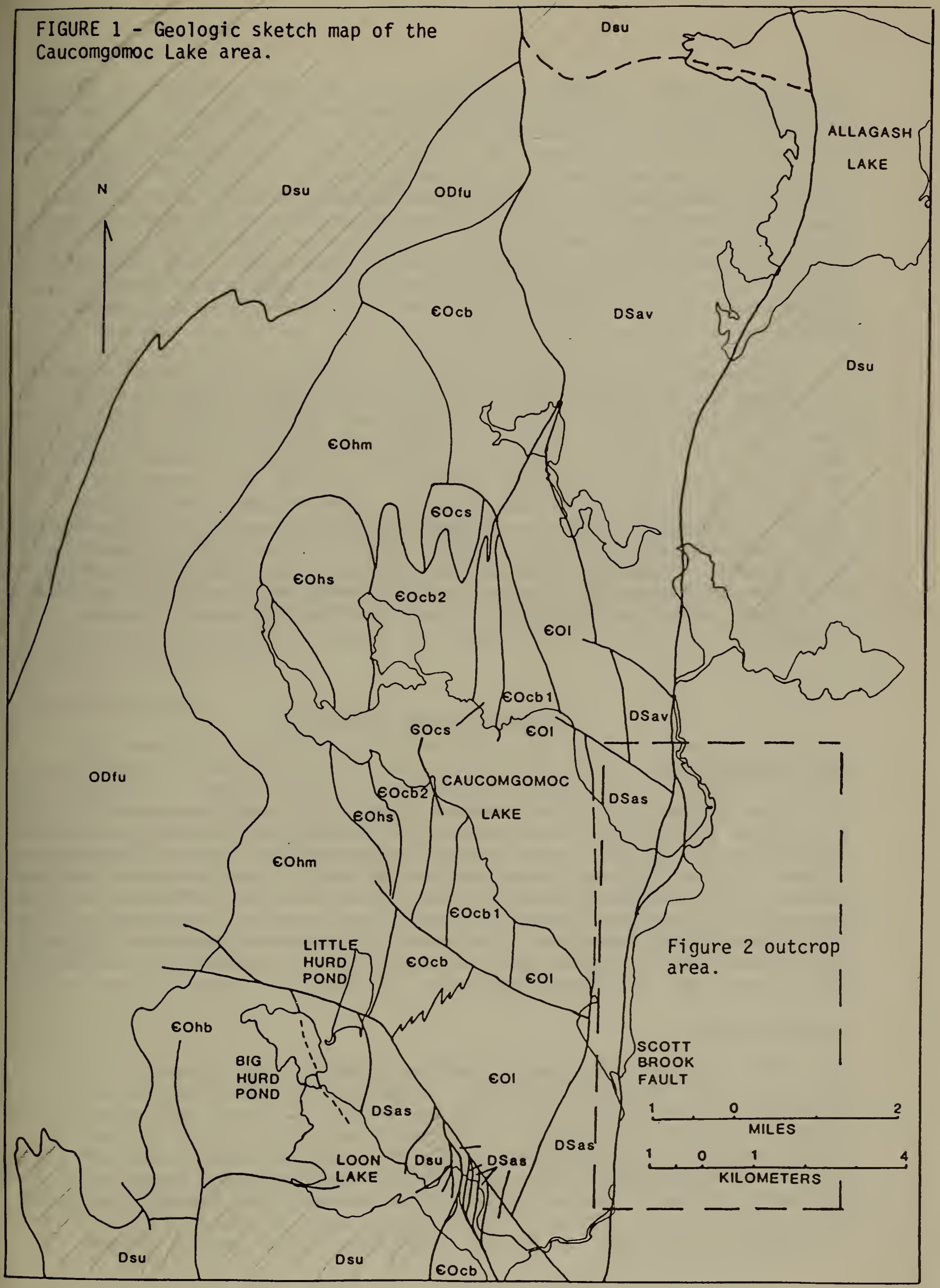
$€Ocb_1$ - lower member on Caucomgomoc Lake. Basaltic lapilli tuffs, basaltic agglomerates, basaltic flows common.

This unit is locally intruded by gabbroic and ultramafic stocks and dikes.

UNIT 1 (Loon Stream Formation)

€Ol - greenish meta-pelite including siltstone, slate and phyllite. Thin to thick bedded ubiquitous laminae and grayish red color common locally.

FIGURE 1 - Geologic sketch map of the Caucomgomoc Lake area.



quadrangle and assigned an Upper Silurian-Lower Devonian age to the sedimentary rocks of the Allagash Lake Formation, while Raabe (1973), considered them to be entirely of Silurian age. Conodonts recovered from limestones of the Allagash Lake Formation during this study are uppermost Silurian (Pridoli) and Lower most Devonian (Gedinnian). (W. Forbes, verbal communication, 1983). The Allagash Lake Formation is correlated with the Spider Lake Formation and East Branch Group which crops out to the northeast in the Munsungun Anticlinorium and several named and unnamed unite in the Moose River Synclinorium (Boucot, 1969), Weeksboro-Lunkunsoos Lake Anticline and Chesuncook Lake Area. (Hall, 1970).

The age of the Seboomook was established as Lower Devonian (Siegenian or Oriskany-Becraft) by Boucot (1961, 1969), in the Moosehead Lake area; Hall, Pollock, and Dolan (1976), in the Matagamon Lake area; and Hall (1970) in the Munsungun Anticlinorium. In northern Aroostook County fossils collected by Boudette, Hatch and Harwood (1976), from the Seboomook are also of Lower Devonian (Oriskany-Becraft, Siegenian) age. Micro fossils recovered from the Seboomook Formation in the Caucomgomoc Lake quadrangle during this study are also of Siegenian age. (W. Forbes, verbal communication, 1983). The Seboomook correlates with the Temiscouata and St. Juste Formations in Quebec and the Littleton and Gile Mountain Formations in New Hampshire and Vermont.

ALLAGASH LAKE FORMATION

Field Relations

The formation is in fault contact with the Seboomook Formation to the east and north and the Pre-Silurian to the west. It is also in fault contact with the Pre-Silurian and Seboomook Formations in a complicated series of small fault slices to the south. The stratigraphic top and bottom of the formation is not exposed or preserved, hence original thickness is unknown. Also, structural complications, poor outcrop, lack of critical structures such as flow tops and bedding planes or topping indicators, make preserved thickness estimates difficult.

Igneous Rocks

The igneous rocks are basalts (chemical analyses in progress). They are aphyric with intergranular or intersertal textures. Dominant structures are thick pillowed and nonpillowed flows. Pillow structures are approximately 1 m to 2 m long by 0.5 m thick. Chill margins are 2 cm to 5 cm thick. Pillow interstices are strikingly mineralized with epidote and hematite. Flow tops are commonly brecciated. Thickness of the brecciated zones is 1 m to 3(+) m. Breccia fragments range in size from 0.5 cm to 6 cm and are angular in shape. Columnar jointing is common

locally. Vesicles and amygdules are generally uncommon. Where they occur they range in size from 0.25 cm to 1 cm.

Sedimentary Rocks

There is a relatively great diversity of sedimentary rock types and sedimentary structures within the Allagash Lake Formation. The sedimentary rocks range in composition from mature quartz arenites to feldspathic and volcanic arenites and wackes. Texturally, the rocks range from silty mudstone or shale through siltstone and sandstone to pebble and cobble conglomerate. The finer texture range is commonly cemented with hematite or limonite producing red bed or rusty weathering sequences. Locally limestones are present. These contain ostracod, bryozoa, coral, crinoid, and brachiopod faunas. The classification of these limestones is fossiliferous micrite to sparse biomicrite. Detrital grains normally make up less than 15% of these limestones

Sedimentary structures are as diverse as the rock types. Bedding in all composition and texture classifications is variable ranging from a few centimeters to approximately a meter or more thick. Beds are rarely graded and may exhibit low angle cross stratification. A variety of cross stratification types are present including herring bone, trough, and planar sets. Very rarely there is evidence of dessication cracks and raindrop(?) impressions.

Interpretation

Presently the basalts of the Allagash Lake Formation are considered to have erupted into a shallow marine or subaerial environment. Brecciated or columnar jointed flow tops are commonly overlain by red silty mudstone and siltstones. Other sediments associated with the flows suggest deposition as a clastic tidal and shallow subtidal facies. These include sedimentary textures and structures documented by Klein (1977) and Ginsburg (1975), to have occurred in intertidal sand bars, channels and channel bars as well as features suggesting exposure and *in situ* weathering and color genesis of sediment on basalt flow tops, similar to those reported by Farrell and Norton (1978), in the Eastport area.

SEBOOMOOK FORMATION

General

The Seboomook Formation is a major stratigraphic unit which crops out over large areas of northern and western Maine. The term "Seboomook" has come to be a widely accepted and generically

applied name for monotonous, cyclically bedded slates and sandstones of Siegenian (Oriskany-Becraft) age within Maine. Perkins (1925, p. 374), first described the Seboomook as "...a dark bluish slate often with well developed cleavage..." and "...sandstone layers (with alternate layers of light and dark slate)...". Perkins derived the name Seboomook slate from Seboomook Dam. He did not establish a type section but rather referred to two localities; one for the "best exposures" of the sandstones at Seboomook Dam, and one for "typical slate" two miles south of the "Y" on the Rockwood-Seboomook Road. Boucot (1961), redefined the Seboomook slate as the Seboomook Formation with a type section "at the east end of Seboomook Lake...and on the Penobscot River for a mile downstream." Boucot's lithologic description (p. 170), of the Seboomook, is as follows:

> "The Seboomook Formation consists almost entirely of cylically layered dark sandstone and slate. The cyclic layering resembles varves and the sandstone layers grade upwards into the slate layers... The layers are fractions of an inch to several feet thick..."

Boucot (1969, p. 34), discussed the variety and proportions of slate and sandstone lithologies in the Moose River Synclinorium. Slate included greater than 95% in certain areas while sandstone at the type locality was more than 50% of the unit. Similarly, Hall (1970, p. 41), considers sandstone to be less than 50% of the formation. Boudette, Hatch and Harwood (1976), consider the Seboomook to consist of variable proportions of calcareous graywacke, gray slate and cyclically bedded gray slate. This usage was followed by Roy (1980). Boudette, Hatch and Harwood (1976), and Roy (1980), were able to subdivide the Seboomook into a lower graywacke and gray slate, or gray slate containing minor graywacke phase and an upper cyclically bedded gray slate and sandstone phase.

The stratigraphic base of the Seboomook varies locally. The stratigraphic top is not well established in the central and western outcrop areas. The total preserved thickness appears to increase to the west and north. The Seboomook is considered by Boucot (1961, 1969), to rest unconformably on granitic and gneissose basement in western Somerset and Franklin counties, and also on the Beckpond limestone and Silurian Hobbstown and Hardwood Mountain Formations in the Moose River Synclinorium. The Seboomook is also unconformable on Lower Paleozoic (Cambrian through Silurian), units in the Matagamon Lake area and Munsungun Anticlinorium. At the western margin of the Moose River Synclinorium Boucot (1961), considers the contact of the Seboomook with the underlying Frontenac Formation to be gradational. Pollock (1983), also suggests the possibility of a conformable contact of the Seboomook with the underlying Frontenac in the western portion of

the Caucomgomoc Lake and Allagash Lake quadrangles. In northwestern Aroostook County, Boudette, Hatch and Harwood (1976), and Roy (1980), show the Seboomook to be in contact with the Silurian "Five Mile Brook sequence" and Lower Devonian "Hafey Mountain sequence." Roy (1980, p. A-24), suggests a possible tectonic (shear zone) contact between the Seboomook and "Five Mile Brook sequence."

The upper contact of the Seboomook is gradational and conformable with the Matagamon sandstone (Pollock, 1972; Hall, Pollock and Dolan, 1976); in the Matagamon Lake area and with the Tarratine Formation (Boucot, 1961, 1969), in the Moose River Synclinorium. The stratigraphic top is not established in the central and western portions of the outcrop belt. Total thickness of the Seboomook varies. Neuman and Rankin (1966), estimate approximately 1,200 m in the Matagamon Lake area. Boucot (1961), estimates 3,000-6,000 m in the Moose River Synclinorium and Boudette, Hatch and Harwood (1976), estimate 5,000 m in northwestern Aroostook County. Overall there appears to be stratigraphic thickening westward.

The internal stratigraphy of the Seboomook is not well established. Overall there appears to be a sandier section near the base as reported by Hall (1970); Boudette, Hatch and Harwood (1976); and Roy (1980). Boucot (1969), established the Camera Hill greenstone member in the upper portion of the Seboomook and the Bear Pond limestone member of uncertain stratigraphic position There is a great deal of room for careful and detailed work in the subject area of Seboomook stratigraphy.

LITHOFACIES

Five lithofacies have been recognized in the Caucomgomoc Lake and Allagash Lake quadrangles. These lithofacies are similar to, or idential to lithofacies both mapped and unmapped for the Seboomook. Criteria used to differentiate the five lithofacies include gross rock type (i.e., slate, wacke, conglomerate) bedding characteristics and sedimentary structures. The five lithofacies include: 1) conglomerate; 2) medium to thick bedded (15 cm to 1 m) wackes with thin (less than 10 cm) claystone or siltstone slate interbeds; 3) thin (less than 10 cm) bedded wackes with thin (less than 10 cm) claystone slate interbeds; 4) thin bedded claystone slates with abundant laminae of silt or sand sized wacke; and 5) medium to thick bedded claystone slate with rare parallel laminae of silt sized material.

1. Conglomerate - Conglomerates form a minor but mappable lithology within the Seboomook Formation east of Caucomgomoc Lake. The conglomerate is lithogically similar to conglomerates of the Allagash Lake Formation. These consist of pebble to cobble sized clasts of angular to subrounded aphanitic

volcanics. The conglomerate is clast supported. Bedding is indistinct and crude grading is observed locally. Contacts of the conglomerate and the other lithofacies of the Seboomook are not exposed. These are interpreted as a submarine channel fill. Hall, Pollock and Dolan (1976); and Hall and Stanley (1972); recognize submarine channels in the Seboomook to the east in the Matagamon Lake area.

2. Medium to thick bedded wackes with thin claystone or siltstone slate interbeds - This lithofacies consists of poorly sorted fine to very fine grained quartzose and feldspathic wacke. Beds range from 15 cm to 2 m in thickness. The bases are sharp. Sole markings are rare. Claystone or siltstone beds are thin (less than 10 cm). Interpreted mechanism of deposition is by turbidity current. Bouma Sequences (Bouma, 1962), include only rare graded or massive intervals. Most commonly the sequences are "bc" and "bcd". The c interval consists of current ripple, rare climbing ripple and convolute laminae. Burrow structures in the "d" interval are uncommon. The turbidites commonly have repetitive bc sequences. These are interpreted as possible submarine fan or fan channel assemblages. Sequences of this lithofacies are not common.

3. Thin (less than 10 cm) bedded wackes with thin (less than 10 cm) bedded claystone slates - Wackes of this lithofacies consist of poorly sorted, very fine sand to coarse silt in chlorite and sericite matrix. Wacke beds average 6 cm in thickness; sole markings are very rare to absent. Sedimentary structures are most commonly the bc turbidite sequence. The c interval is ripple cross lamination. The claystone slate appears structureless, but may be texturally graded. Burrows are rare. Sand : slate ratio of this lithofacies is approximately 1 : 1. The wackes of this lithofacies are interpreted as "distal" turbidites relative to the wackes of the preceding lithofacies. Deposition of this lithofacies is interpreted to have occurred on the lower or distal portion of submarine fans. This lithofacies is moderately common.

4. Thin bedded claystone slate with abundant laminae of silt or sand sized wacke - The claystone slate beds commonly range from 1 cm to 10 cm in thickness. Textural gradation in these beds may be present. Burrows are rarely present. The laminae are commonly sharp and texturally grade upwards into the overlying claystone. Ripple lamination is common within these laminae. Average thickness of these laminae is 5 mm and they number from approximately 10 to 50 or more per meter. This lithofacies is interpreted to have been deposited as hemipelagic sediment and/or mud turbidites. Similar lithofacies and interpretation is recorded in the Seboomook by Hall and Stanley (1973). This lithofacies is very common in the southern and eastern portions of the area.

5. Medium to thick bedded claystone slate with rare parallel laminae of silt sized material - The claystone slate beds commonly range from 15 to 50 cm in thickness. Bedding planes are indistinct and where seen are delineated by very fine silt laminae. This lithofacies is characterized by "paper thin" cleavage. Parallel laminations of very fine grained silt are the only sedimentary structures seen. Deposition is interpreted to have been in the hemipelagic or pelagic environment. This lithofacies is most common in the western outcrop area.

DEPOSITIONAL ENVIRONMENT

Depositional environment of the Seboomook has been interpreted by Hall, Pollock and Dolan (1976), to have been a pro-delta marine slope association, where the Seboomook represents a classical shaly flysch. This interpretation is consistent with the interpretation of depositional mechanism and environments included with the lithofacies as discussed here. It is suggested that the basin margin and slope was to the east. Sediment transport was from east to west (Hall, Pollock and Dolan, 1976, p. 61; and Pollock, 1972), at the basin margin. In the portion of the basin discussed here, deposition was predominantly hemipelagic and pelagic with submarine fan deposits. Sediment transport in this outcrop area was parallel to the basin axis in a northerly direction (Pollock, unpublished).

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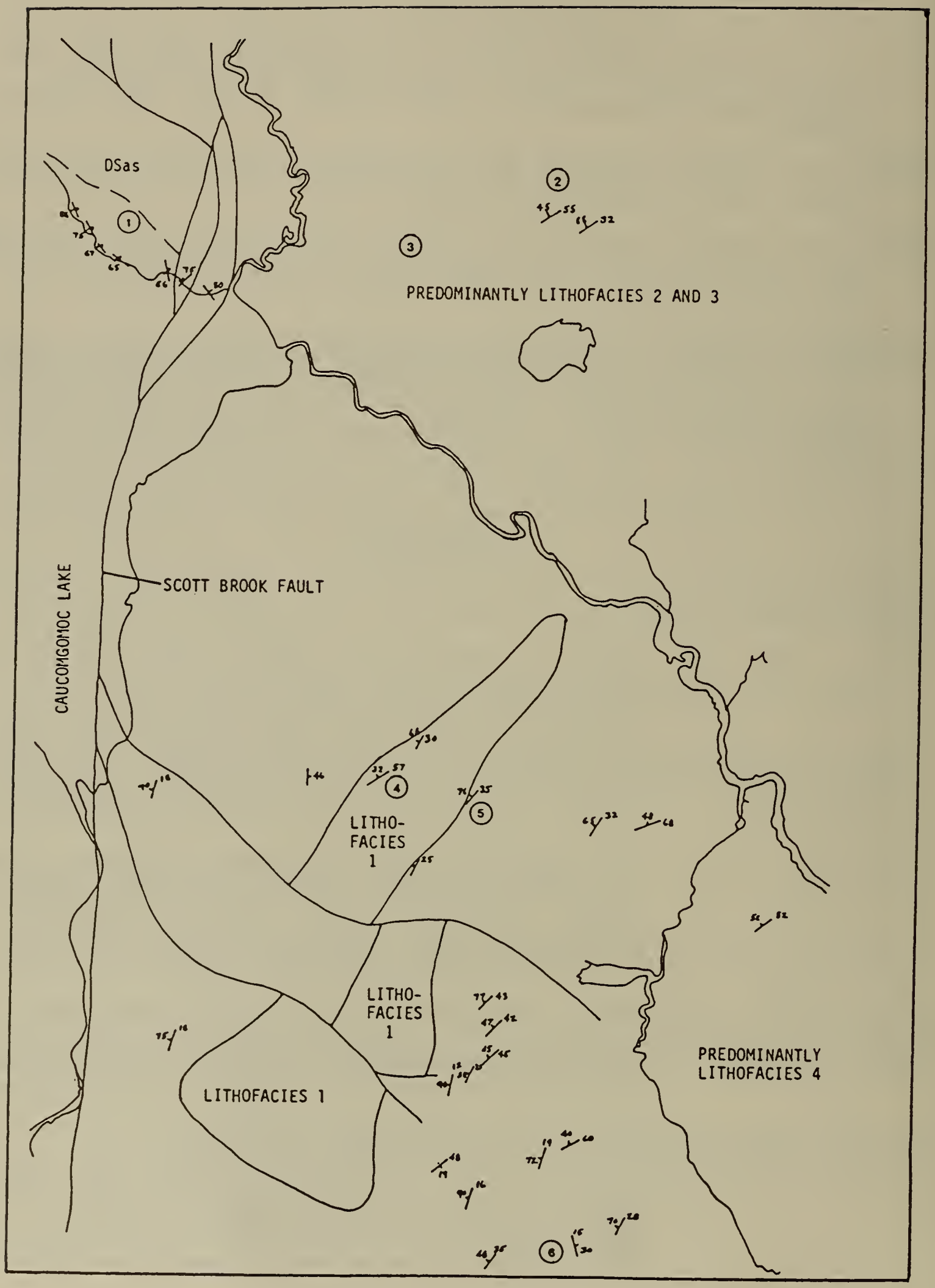
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ITINERARY

TRIP		
0.0	0.0	8:00 a.m. leave from Pray's store and travel west on the Golden Road. Note: This trip involves travel on remote, privately maintained woods roads were access is controlled. Your vehicle should be equipped with good tires, a good spare and a reliable jack. You should also have a full tank of gas at the start. This trip would not have been possible without the permission and courtesy of North Maine Woods, Ashland, Maine; Seven Islands Land Company, Bangor, Maine; and Great Northern Paper Company, Millinocket, Maine.
7.8	7.8	Road leading to Greenville is on the left.
17.0	9.2	Outcrops of Seboomook Formation will be seen now for the next several miles (no stopping).
24.5	7.5	West branch Penobscot River
26.2	1.7	Ragmuff Gate. This gate is locked and maintained by Great Northern Paper Company. For the rest of the trip we will be on roads to which there is no general access by the public.


DSas
1
2
3
PREDOMINANTLY LITHOFACIES 2 AND 3
SCOTT BROOK FAULT
CAUCOMGOMOC LAKE
4
5
6
LITHO-
FACIES
1
LITHO-
FACIES
1
LITHOFACIES 1
PREDOMINANTLY
LITHOFACIES 4

		Speeds on these roads will be between 25 and 45 mph. Do not deviate from the route described or lag behind. The terrain for the next several miles is relatively flat. Outcrop is sparse. This is typical of terrain underlain by the Seboomook Formation.
32.8	6.6	Former site of Great Northern's Ragmuff logging camp.
40.6	7.8	Bear to left at Forestry sign.
40.9	0.3	Junction of Four Roads. Bear to right on the road which is maintained. (Not a sharp right onto the road with the grassy center leaving the gravel road on your left.)
41.2	0.3	Town line marker.
44.3	3.1	Locked gate. We are now leaving Great Northern lands. Lands that we are now driving on are managed by Seven Islands Land Co. Continuing down this road, ridges which appear before us are the outcrop area of the Allagash Lake Formation.
46.3	2.0	Caucomgomoc Lake Dam. From the dam you have a panoramic view of Caucomgomoc Lake. Low hills to the north (right, as you face Caucomgomoc Lake), are the outcrop area of Allagash Lake Formation. The hills in the distance to the west (looking down the lake), is the outcrop area of the Frontenac. The lower hills, also to the west and closer to the lake, are Pre-Silurian outcrop. The trace of the Scott Brook fault trends approximately north-south between the dam and the Allagash Lake Formation. Scott Brook, Loon Stream and Ciss Stream all follow this fault. The Scott Brook fault separates the Allagash Lake Formation from the Seboomook Formation. Cross the dam and proceed.
47.9	1.6	LEFT TURN
50.9	3.0	Ciss Stream Bridge. (As this itinerary goes to press, the bridge is badly in need of repair. Hopefully it will still be passable for the trip.)

51.2	0.3	Junction of logging road with small, unmaintained spur. The party should combine themselves now into the highest clearance vehicles. We will proceed down the small spur road approximately two miles to the shore of Caucomgomoc Lake. This road <u>should</u> be passable to two-wheel drive vehicles. There may be a few wet and muddy spots and care should be taken to navigate them correctly. STOP 1. Proceed to the shore of Caucomgomoc Lake and traverse westerly along the shore approximately 0.6 mi. This section, at times of low water, best illustrates the variability of sedimentary rock types and sedimentary structures in the Allagash Lake Formation. The rocks include redbeds, quartz arenites and limestones with diverse fauna. The section is overturned and small folds may be observed locally. The interpretation put forth is that these rocks were deposited in shallow water, perhaps shallow subtidal and tidal environments. Comments and criticisms would be greatly appreciated. Return to vehicles and retrace route to the "main" road. We will begin to retrace our route. The remainder of stops today will be in the Seboomook Formation.
51.5	0.3	Ciss Stream Bridge
51.5	0.6	This is a newly constructed and active "borrow pit". The Seboomook here is typically a hybrid between lithofacies 4 and lithofacies 5. The slate here is uniformly and ubiquitously laminated with silt sized material.
53.8	2.3	Logging camp. Continue straight.
54.6	0.8	STOP 2. Outcrops are on either side of the left hand spur road. The best outcrop is to your left as you proceed down the road. The section here is representative of lithofacies 3. The Seboomook consists of thin bedded ripple cross-laminated wacke with thin interbeds of claystone or siltstone slate. Return to vehicles and retrace route.
55.4	0.8	LEFT TURN

55.8	0.4	STOP 3. (Optional) Outcrop is a pavement in a logged over yard on the right. The section consists of lithofacies 3, and is very similar to STOP 2.
57.0	1.2	Caucomgomoc Dam
59.5	2.0	Great Northern Paper Company gate.
59.8	0.3	LEFT TURN
60.3	0.5	STOP 4. Conglomerate lithofacies. In this immediate area are several small knolls of pebble conglomerate. The clasts are commonly felsic to intermediate aphanitic volcanics. Shades of gray are the most common color of the clasts, but red and green coloration may be seen. The clasts are for the most part rounded and textural gradation may be seen in some beds. Bedding is commonly thick and bedding planes are indistinct. Trough cross beds are locally present, the major structure is the regional cleavage and small en echelon quartz veins. Proceed.
60.8	0.5	LEFT TURN and proceed without turning.
61.9	0.7 to 1.1	STOP 5. This stop illustrates the nature of sedimentary slumping that is locally present in the Seboomook. Outcrops are present at the intersection of two roads and down the left hand spur road. Retrace route to the Caucomgomoc Dam road.
63.6	1.7	LEFT TURN
66.8	3.2	LEFT TURN. Proceed down road with grass growing in the center strip. (Road to Black Pond)
67.4	0.6	Town line post on right.
67.5	0.1	LEFT TURN
68.0	0.5	RIGHT TURN
68.5	0.3 to 0.5	STOP 6. There are several small outcrops to the north (left). This is the thin bedded claystone slate with abundant laminae of silt or sand sized wacke lithofacies.

		This is the most common lithofacies of the Seboomook in the area. The outcrops here are typical of size and extent of this particular lithofacies. You may observe bedding, determine stratigraphic tops, observe refracted and kinked cleavages, and see local minor folds here. Retrace route for approximately 1.3 miles.
69.8	1.3	LEFT TURN
70.1	0.3	Junction of four(4) roads. Continue straight.
71.3	1.2	Intersection with Loon Lake Road on right. Continue straight.
74.6	3.3	Site of old Scott Brook logging camp. Continue straight. DO NOT TURN LEFT.
77.2	2.6	Small Stream. Continue straight.
77.3	0.1	LEFT TURN
78.3	1.0	Outcrops of lithofacies 2 on right.
78.7	0.4	STOP 8. (Location is south of the sketch map.) Outcrop is on the right and in the center of the logging road here, and about 0.1 mile further down the road. Outcrops consist of lithofacies 2. This area is one of the most accessible and prominent areas of this lithofacies. You will have an opportunity to observe sedimentary structures, bedding characteristics and rock types. This concludes the excursion. Return to vehicles and retrace route back to the "main" road.
80.1	1.4	RIGHT TURN
87.3	7.2	Sharp right turn at four road intersection.
101.8	14.5	Great Northern Paper's Ragmuff Gate. LEFT TURN
120.2	18.4	Right turn for those of you who are going the 35 miles to Greenville. Straight for those of you who are going to Millinocket.
128.0	7.8	Pray's Store. End of trip.

TRIP B-4

THE TIMING OF THE ALPINE GLACIATION OF MT. KATAHDIN

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INTRODUCTION

Mt. Katahdin is the highest mountain in Maine (1605m) and, with a local relief of about 1450 meters, is one of the largest mountains east of the Rocky Mountains. Griscom (1966) provides an excellent summary of the early ascents and geologic exploration of Mt. Katahdin. The mountain is composed of granite and is part of a large Devonian pluton that intrudes lower and middle Paleozoic sedimentary and volcanic rocks. Mt. Katahdin is part of a large massif in the northernmost 15 percent of the area of the whole pluton. The rest of the pluton consists of rolling hills of no more than 200m relief. Hon (1980) points out that most of the pluton consists of coarse granite without an interlocking texture which readily disintergrates to grus. Hon goes on to say that only where the main granite body is covered by a fine grained summit facies rock with a tough, interlocking texture do mountain occur within the Katahdin pluton.

Erratics found by Tarr (1900) and Antevs (1932) near the summit of Mt. Katahdin and by Caldwell (1972) on other mountains in the region support the view that Mt. Katahdin was covered by continental ice sometime in the Pleistocene. There is no direct evidence that the highest elevations were covered by Late Wisconsin ice, although theoretical ice profiles constructed by Davis (1978) suggests that they were.

The purpose of this field trip is to review evidence for and against late Wisconsin Alpine Glaciation on Mt. Katahdin. Caldwell (1959, 1966, 1972, 1980, 1982) has long held that Alpine Glaciation did occur within the late Wisconsin. Davis (1976, 1978, 1983; Davis and Davis, 1980; and Davis and others 1980) has proposed that there is no concrete evidence for late Wisconsin Alpine Glaciation. In this field guide ideas held by Caldwell will be so identified while the views of Davis will also be indicated.

FEATURES FORMED BY ALPINE GLACIATION

Cirques

In the Mt. Katahdin area cirques occur only on those mountains underlain by Katahdin granite. Most of the cirques (6-7) were formed on the massif which includes Mt. Katahdin and the other high peaks which lie above timberline. The 3 largest cirques are on the east side of the mountain and have headwall heights which range from 720 meters to about 100 meters. These 3 great cirques have flat to concave floors and steep headwalls composed largely of bedrock. Postglacial rockfall and avalanche debris does mask the lower slopes of the cirque headwalls and sidewalls. The total aspect of these cirques is one of remarkable freshness, especially when compared with other cirques in northeastern United States assumed to be occupied by glaciers in the late Wisconsin (Wagner, 1970; Craft, 1979; Bradley, 1981).

Aretes

The 3 east-facing cirques are separated by aretes. The most typical arete is Hamlin Ridge, which separates North Basin cirque from Great Basin cirque (Figure 1). The arete which separates Great Basin cirque and South Basin cirque, Cathedral Ridge, has been shortened and lowered by glacial erosion and mass wasting. The most spectacular serrate mountain crest is the Knife Edge, but it may not technically be an arete because there is no cirque on its south side. However, the long narrow saw tooth ridge crest and the over 2000 foot (720m) drop into South Basin more than make up for this deficiency.

Moraines

The aspect of alpine glaciation on Mt. Katahdin about which there is the greatest controversy concerns moraines found within and down the mountain from the three largest cirques. Tarr, 1900; Antevs. 1932; and Caldwell, 1959, 1966, 1972 and 1980; identified moraines in each of these cirques. In addition these authors believed the large Basin Ponds moraine was a medial moraine formed between the combined alpine glaciers from the three cirques and the still active tongue of the Laurentide ice sheet. The common interpretation was the alpine glaciers which formed these moraines were both contemporaneous with (at the Basin Ponds Moraine), and postdated (at the moraines with the cirques), the Late Wisconsin ice sheet. Davis (1976, 1978, 1983) believes that the Laurentide ice sheet covered Mt. Katahdin during the Late Wisconsin but that no alpine glaciers postdated the ice sheet glaciation. Davis does not believe that there are moraines within the large cirques and interprets the Basin Ponds Moraine to be a lateral rather than a medial moraine. Davis and Davis (1980) and Davis and others (1980) suggest a piping mechanism to explain the lengthy lag times between deglaciation and accumulation of datable organic material on alpine bog and pond bottoms.

Davis views a glacial sequence for Mt. Katahdin similar to that proposed by Goldthwait (1970) for the Presidential Range in New Hampshire. The features discussed here are located in Figure 1 which also shows the location of field trip stops.

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Figure 1. Vertical aerial photograph made from a false-color image, courtesy of Baxter State Park Authority. North is at top of photo. Scale is about 1cm = 1.3km. Numbers refer to locations of field trip stops. Letters refer to features discussed in text: A. Terrace-like features on south side of the mountain, B. Basin Ponds Moraines, C. Baxter Peak, elevation 1605m.

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ITINERARY

Road Mileage	
0.0 Stop 1	The Katahdin quadrangle covers the area of this field trip. Field trip will assemble at 8:00 A.M. at Abol Crossing, at west end of bridge over West Branch Penobscot River. Park in gravel pit on east end of bridge and walk to west end of bridge. From this point we can view terrace-like features on south side of mountain. Davis believes these are lateral moraines, Caldwell believes they are Kame terraces. Evidence for both points of view will be discussed. In foreground is part of an area (3000 acres) burned during 1976 forest fire. Collapsed outwash and ice contact deposits lie below prominent ridge.
	Return to cars where brief discussion of the rock types found in the drift in this area. These play a large role in today's discussions.
0.1	Bear left on State road.
3.6	Turn left onto Baxter State Park road.
4.8	Togue Pond Camps. A basal radiocarbon date of 11,630 $\pm$ 260 yrs. B.P. (SI-2992) from Lower Togue Pond (east side of road) provides a minimum age for deglaciation of this part of the lowland surrounding Mt. Katahdin (Davis and Davis, 1980)
5.1	Park entrance. No pets, guns, shovels or pick axes allowed in park.
5.2	Turn right on Roaring Brook Road - road to left is built on a large esker that extends back to Abol Crossing.
7.0	Rum Brook.
10.2	Windey Pitch. Caldwell (1980) feels that this ridge is an end moraine because there are no bedrock exposures. However, Davis (unpubl. data) interprets this feature as a large mound of hummocky ground moraine, probably overlying a bedrock knoll.
11.0	Avalanche Brook and Avalanche Field.
13.6	Roaring Brook Campsite and Roadhead. Bear left and park in designated area. Secure cars.

Trail Mileage	
0.0	Trail log from Roaring Brook. Approximate mileage and hiking times do not include time at stops.
0.05	Bear right over bridge onto Sandy Stream Pond Trail.

0.6 (15 Mins.)

Stop 2 Sandy Stream Pond. From this vantage we can compare the aspects of the Basin Pond Moraine with the features seen at Abol Crossing. Also we can see into the large east-facing cirques, should the weather deteriorate before we get there. We also may see a Maine moose or two feeding. Return to Roaring Brook.

1.1 (30 Min.)

Turn right onto Chimney Pond Trail

2.9

Stop 3 Outlook to right of Roaring Brook Trail about 100 meters beyond Halfway Rock (half way between Roaring Brook and Chimney Pond campsites). The sandy drift exposed at the Outlook has a noticeable content of erratic pebbles (40-60%, Davis, 1976). The small ridge dams a small bog with a basal radiocarbon age of 7070 ± 90 yrs. B. P. (SI-1049; Davis and Davis, 1980). This basal date and other radiocarbon dates from lake-bottom sediments on Mt. Katahdin are believed to be 5000 to 9000 years more recent that the time of deglaciation for reasons that will be discussed at Stop 9, Dry Pond. Davis (1976) believes that the small ridge at the Outlook is a recessional moraine formed by a shrinking mass of ice located in the lowlands south and east of Mt. Katahdin. However, Caldwell feels that the ridge could have been formed by cirque ice that overran the Basin Ponds Moraine just above (west) of here. Figure 1 shows location of Stop 3 and other stops which are identified by number on the photograph.

3.1 (1 Hr.-30 Min.)

Stop 4. Basin Ponds. The Basin Ponds Moraine (Figure 1, locality 4) dams the Basin Ponds, as well as numerous other smaller ponds on the east side of Mt. Katahdin. Because nearly all boulder size material in the moraine is composed of Katahdin quartz monzonite, Caldwell infers that the Basin Ponds Moraine was deposited by ice issuing from the 3 cirques to the west and overriding the continental ice here. However, Davis (1976) identified pebbles (2-5 cm, median axis), finding erratic compositions up to 28%. Moreover, the Basin Ponds Moraine extends both north and south well beyond the mouths of the 3 cirques, wraps tightly around Keep Ridge to the south (visible from Stop 3, Blueberry Knoll), and does not drop in elevation north to south, but remains within 2400- to 2500- feet contours. Taken together with the lateral moraines on the south slope of Mt. Katahdin, and up to 6% content of erratic red shale pebbles found in both moraines but not elsewhere on Mt. Katahdin, Davis (1976) infers that the Basin Ponds Moraine was formed by an ice advance from the lowlands to the southeast.

The red slate to which Tom refers is the Capens formation, a Silurian red and green slate, that occurs in the Frost Pond area, 18 miles nearly due west of where we were and in the Allagash Lake area, more than 50 miles northwest. One would want to find evidence other than red slate before posulating an advance from the southeast, where no known outcrops of red slate occur (Caldwell).

A basal radiocarbon date of 5665 ± 110 yrs. B. P. (I-7348) from Lower Basin Pond probably is at least 7000 years more recent than the time of deglaciation.

3.4 (Hr. 45 Min.) North Basin Cutoff Trail.

4.1 Stop 5. Blueberry Knoll and North Basin cirque. Blueberry Knoll is at the mouth of a cirque with the least vegetated floor of any cirque in New England (Figure 1, locality 5). Caldwell believes that Blueberry Knoll is an end moraine deposited by a cirque glacier; however, Davis feels that the shape of the knoll and seismic data suggest a bedrock origin. Davis (1976) found up to 15% content of erratic pebbles on Blueberry Knoll, suggesting meltout from continental ice. Blueberry Knoll not only offers fine views of North and South Basin cirques, but also an excellent view of the Basin Ponds Moraine and associated hummocky topograph to the east and south.

Proceed along trail to middle of North Basin Cirque.

4.4 (3 Hrs.) Stop 6. North Basin Cirque. Two small ponds occur in prominent basin with ridge of bouldery moraine (Caldwell) beyond, near head of cirque is an irregular hommocky mound of debris that Caldwell (1959, 1972) described as a pro-talus rampart but may be a rock glacier. Return to Blueberry Knoll.

4.7 (3 Hrs.- 30 Mins.) Blueberry Knoll. Return to trail.

5.0 (4 Hrs.) Hamlin Peak Trail up Hamlin Ridge. Although hiking up this ridge to Hamlin Peak would provide spectacular views of features under discussion, there will not be time and still have the last two stops. It should be realized that it would take about 3 1/2 hours to reach Hamlin Peak and return to Chimney Pond and it will be getting dark by then.

5.6 (4 Hrs. 30 Mins.) Stop 8. Chimney Pond. A classic tarn provides a spectacular view of the 2000- ft. cirque headwall of South Basin. A basal radiocarbon date of 3050 ± 90 yrs. B. P. (I-7347) is believed at lest 9000 years more recent than the time of deglaciation of South Basin cirque. Although low percentages of erratic pebbles occur in Great and South Basins, large erratic cobbles are present.

6.1 (4 Hrs. 45 Mins) Stop 9. Dry Pond. This depression of boulders holds water only for a few days during spring snowmelt or after intense rainstorms. Fine material is flushed out through cavities among the large boulders on its floor. A similar flushing probably occured during early stages of formation of other bogs and ponds on Mt. Katahdin, thus explaining the extremely minimum-limiting nature of basal radiocarbon dates. Caldwell interprets Dry Pond as a

depression behind a cirque glacier end moraine; however, Davis did not find a ridge in the forest suggestive of a moraine. Numerous large erratic cobbles are found in trail cuts between here and Basin Ponds.

9.4 (6 Hrs.) Roaring Brook campsite and end of trip.

- Thank you -

TRIP B-5

The Northwest Boundary Fault of the Boundary Mountain Anticlinorium

by

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Introduction

The Northwest Boundary Fault (NWBF) has been mapped in detail from north of Little Turner Pond in the Attean 15' quadrangle southwestward to the Maine/Quebec border, a distance of 12 miles (Figure 1). This work was done during a four week period in 1978 and, due to the very limited access and time, mapping was generally confined to within one half mile of the fault. Work was supported by the Maine Geological Survey and the Nuclear Regulatory Commission.

Geologic Setting

One of the prominent and apparently unique tectonic terranes in New England is the Chain Lakes massif of Boone and others (1970) which cores the NE-SW trending Boundary Mountain Anticlinorium (Albee, 1961). Some rocks within the massif have been dated by Naylor and others (1973) at 1.5 b.y. The northeastern end of the massif was intruded in the Upper Middle Ordovician by the Attean Quartz Monzonite (Albee and Boudette, 1972), and these two units constitute the pre-Upper Silurian rocks along the southeast side of the NWBF. The post-Upper Silurian rocks on the northwest side of the NWBF include slivers of Upper Silurian limestone and Lower Devonian phyllitic slates of the Seboomook Formation. Within the fault zone, a wide variety of breccias have been observed.

Stratigraphy

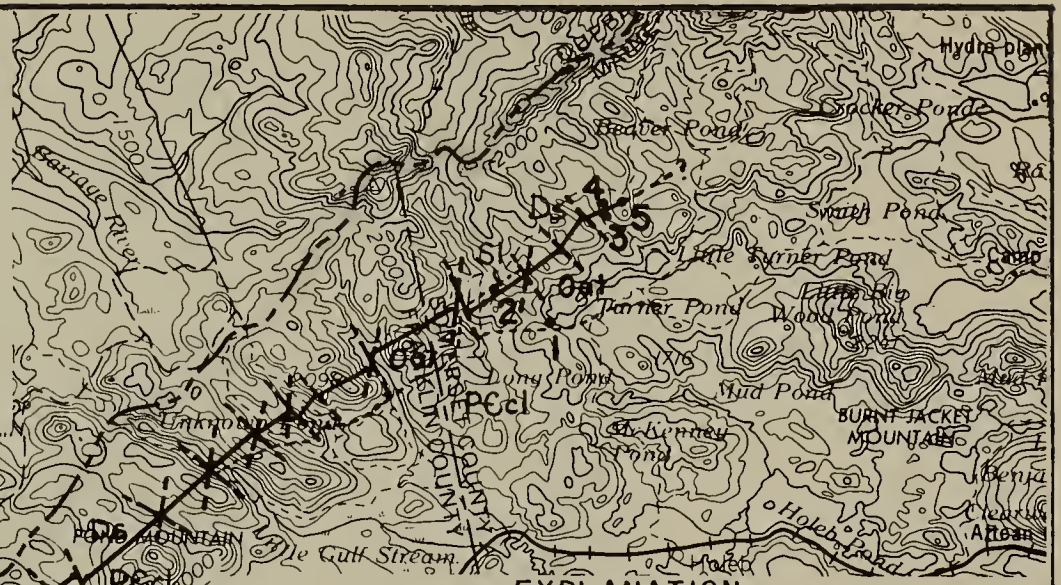
The following lithologic descriptions refer only to those portions of named units which crop out along the trace of the fault. All rock names are based on hand specimen identification in the field.

Chain Lakes massif (Precambrian) - Rocks of this unit locally consist of a wide variety of mafic and felsic metavolcanics. The mafic rocks are characteristically dark green and often have a purplish tint. They vary in grain size from very fine grained to medium grained, and their textures range from strongly schistose to massive. Individual rock types include metamorphosed crystal tuff with euhedral feldspar and minor quartz, metabasalt and metadiabase (both greenstones), pyroxene-plagioclase granofels, albite-chlorite-quartz schists, and fine-grained feldspar-chlorite-biotite-quartz gneisses with "eyes" of quartz.

Felsic metavolcanics include pale yellow fine-grained siliceous rocks, whitish gray cherty rhyolite (?), quartz-limonite-muscovite schist, pale pink siliceous muscovite schist, brownish gray feldspathic granofels, pale greenish gray siliceous schist, yellowish pink quartz-porphory latite. Grain sizes are generally very fine to fine, and only rarely medium. Most of these rocks have well-developed schistocity parallel to close-spaced brittle fractures, although the cherty rhyolite (?) is massive. As a general rule, the felsic rocks are located between the mafic rocks and the NWBF.

GEOLOGIC STRUCTURE MAP OF THE NORTHWESTERN PORTION OF THE "CHAIN LAKES MASSIF" AND VICINITY





EXPLANATION

Ds	Seboomook Formation
Sl	Silurian limestone
PCcl	Chain Lakes massif
Dsp	Spider Lake Granite
Dbi	Big Island Pond Pluton
Dcp I + II	Chain of Ponds Pluton I = Coarse border zone II = Porphyritic interior
Oat	Attean Quartz Monzonite
x	Mt. Louise

N

0 5 10 15 km

0 5 10 miles

SCALE 1:250,000

Figure 1

Attean Quartz Monzonite (Upper Middle Ordovician) - The Attean Quartz Monzonite occurs along the southern side of the contact in the Attean and Skinner Northeast Quadrangles, and in most of the Skinner Quadrangle. It is an extensively mylonitized, two-feldspar granite in most locations, commonly porphyritic with varying amounts of recognizable potassium feldspar phenocrysts. The percentage of quartz appears to range from as low as 20% up to 35%, but mylonitization makes these estimates difficult.

The texture of this unit varies with proximity to the contact, becoming finer grained to the northwest where the extent of mylonitization is greater. The rocks commonly contain oriented chlorite, producing a foliation and a pale green color on fresh surfaces. Outcrops in the Attean Quadrangle were observed to contain secondary calcite as well as chlorite. The weathered surfaces of the Attean Quartz Monzonite are typically whitish gray to strongly pink with increased extent of shearing and recrystallization.

Metalimestone (Upper Silurian) - As noted by Albee and Boudette (1972), thin slivers of Silurian limestone are found in the Attean Quadrangle along the contact between the pre-Upper Silurian rocks and the Devonian slates. One such sliver is well exposed in the Skinner Northeast Quadrangle east of East Branch Gulf Stream (Stop 2). The rocks are exposed over a distance of 610 m with a map width ranging from 61 m at the west end to 92 m at the east end of the outcrop belt (see Figure 2).

These limestones are typically gray on fresh surfaces, weathering to brown. The most common texture of the limestone is massive, but exposures of thinly laminated (1 mm) limestone were also observed. One outcrop exhibited a swirled texture of very fine-grained gray limestone and brownish weathering coarser limestone with disseminated black angular fragments. Mylonitized breccias of limestone and Attean Quartz Monzonite occur along the southern margin and within the outcrop belt.

West of the limestone sliver discussed above, two occurrences of limestone were observed. The first is in East Branch Gulf Stream where large angular boulders of pale purple, laminated limestone occur, the lamination being seen as discontinous layers of very fine-grained and medium-grained (1-2 mm) limestone. Square and circular particles can be seen in the coarser layers with the circular fragments having darker cores. These particles are thought to be crinoid stem fragments. The only other limestone occurrence in the map area is in the Skinner Quadrangle, between the Devonian slate and the Attean Quartz Monzonite, at the crest of the first major hill west of West Branch Gulf Stream. The limestone at this locality is fine grained and has a brownish gray color, appearing to be very similar to the major sliver described above.

Seboomook Formation (Lower Devonian) - Dark gray phyllitic slates are exposed on the northern side of nearly the entire extent of the contact between the pre- and post-Upper Silurian rocks. In those areas where Silurian limestones constitute the boundary, the Devonian slates are exposed immediately to the north of the limestone.

In the Skinner Northeast Quadrangle, in the eastern portion of the map area, slates are typically laminated on a scale of 1 to 2 mm with black metapelite beds and dark gray metasiltstone beds nearly parallel to a phyllitic slaty cleavage. Discontinuous 1 cm thick lenses of metasandstone occur commonly in places but rarely in most localities. Locally bedding can be observed dipping gently to the NW.

Dark gray phyllitic slates are common in the western portion of the map area, but they are rarely laminated. The grain size of these rocks is generally somewhat coarser (metasiltstone) than is that of the slates to the east, and phyllitic texture commonly is absent. The color of these rocks varies from black to gray, and weathered surfaces typically are brown and rusty.

Breccias (age unknown) - A wide variety of breccias are exposed in various locations along the length of the contact between pre- and post-Upper Silurian rocks, as well as in rocks immediately adjacent to that contact. In the eastern portion of the map area, near the Silurian limestone sliver, breccias of limestone-in-Attean Quartz Monzonite matrix, Attean Quartz Monzonite-in-limestone matrix, limestone-in-Devonian slate (schistose) matrix, and Devonian slate-in-limestone matrix have been observed. Clast sizes commonly reach 10 cm in their long dimension and rarely reach 20 cm. These breccias have angular, elongated clasts showing internal deformation, contained in a schistose matrix. The breccias generally are found at the lithologic contacts, but one limestone-in-black schist breccia occurs within the limestone outcrop belt.

A breccia containing clasts of Attean Quartz Monzonite, limestone, and black schist is exposed on the west shore of Gulf Stream at the contact. This rock is intensely sheared with a well-developed schistose texture and a dark green color.

Near the western outcrop limit of the Attean Quartz Monzonite, two exposures of breccia occur on opposite sides of an unnamed stream west of Gulf Stream. Both of these breccias consist of very angular, white cherty volcanic(?) fragments up to 10 cm in length contained in a granitic matrix. Abundant coarse quartz is present in the matrix along with small angular fragments of the banded cherty volcanic(?) rock. The western exposure of this breccia is approximately 30 m south of typical Devonian slate.

Farther to the west, near the west side of Little Gulf Stream, a breccia of angular white, cherty volcanic(?) fragments in a fine-grained, felsic schistose matrix is exposed. This breccia is very similar to rocks at another exposure located farther to the west, north of the summit of Clear Pond Mountain in the Boundary Pond Quadrangle.

All of the breccias share the characteristics of having clasts elongated parallel to the regional fracture foliation (N20-40°E, 55-80°W) and having extensive phyllosilicate mineralization on their sheared surfaces. These surfaces are commonly lineated, but no movement directions were obtained.

Structural Geology

Features in pre-Upper Silurian rocks - Well-developed, closely-spaced intersecting brittle fractures with superimposed joint sets are the dominant structural features in the pre-Upper Silurian rocks. No faults or folds were mapped within the units of this group, but the lack of planar features would make the recognition of such features unlikely. Strongly mylonitized Attean Quartz Monzonite occurs in topographic depressions which parallel the NWBF, suggesting that faults within those rocks trend parallel to their northwestern contact.

The orientations of the closely-spaced brittle fractures show a consistent trend at a shallow angle to the northern boundary of these rocks. Strikes of these fractures are typically N20-40E, and dips vary from 60 to 85W. These fracture generally occur in sets that intersect at angles of 10 to 20 degrees or in a single set, producing in each case a pervasive fracture foliation. The spacing between

fractures is generally less than 1 cm. Several fine-grained rock types in both the mafic and felsic metavolcanic group commonly have a schistocity that parallels or replaces the fracture foliation. Both the fracture foliation and the schistocity tend to increase in their degree of development as one approaches the NWBF. Fracture foliation is commonly very well developed in the Attean Quartz Monzonite where it is associated with observable mylonitization. Both quartz and feldspar generally exhibit the results of a mechanical reduction in grain size.

Many outcrops exhibit slickensides and mineral lineations on the steeply dipping fracture foliation surfaces. The direction of movement is almost directly down dip (normal), that is N side down, in nearly all cases. Occasionally lineations occur plunging at an angle less than the dip of the fracture surface, and very rarely slickensides indicate high angle reverse motion.

Joints in the pre-Upper Silurian rocks are generally well-developed, nearly vertical sets. They are typically planar and smooth, and their spacing is variable but usually consistent at the outcrop scale (1 cm. to 2 m.). Most joints are devoid of mineralization, but quartz-filled joints are locally abundant. This is particularly true at closest proximities to the NWBF. Rarely, quartz veins roughly parallel to the joint sets account for as much as 40% of an outcrop by volume.

Features in post-Upper Silurian rocks - The dominant structural feature in the Upper Silurian and Lower Devonian rocks is a well-developed slaty cleavage, commonly exhibiting a phyllitic sheen. Where bedding is observable, the cleavage is generally parallel or nearly parallel to the bedding. The shallow angles of intersection produce a nearly horizontal lineation on the cleavage surfaces which usually dip at a slightly steeper angle. A second cleavage is occasionally observed, generally associated with kink folds. The sense of rotation on these kinks is consistently down dip either to the northwest or southeast. A well-developed slaty cleavage occurs in some outcrops of the Upper Silurian limestone trending parallel to the regional strike and dipping moderately to steeply to the northwest. Many outcrops of the limestone are massive and have poor cleavage.

Joints in the post-Upper Silurian rocks are well developed, planar and smooth, and generally occur as one or two sets at high angles to the cleavage. The occurrence of quartz in these joints is most common near the NWBF.

The Northwest Boundary Fault - The topography of the mapped section of the NWBF is characterized by north and northwesterly trending stream valleys which cut across the strike of the fault. Hills between the valleys have an irregular topography with ridges and gullies parallel to the NWBF. Between each pair of adjacent streams the contact between pre- and post-Upper Silurian rocks can be mapped in detail, and the orientation of each segment of the NWBF is tightly controlled. Adjacent segments of the fault may be offset as much as 750 feet and the strike of the fault may change as much as 25 degrees from one segment to the next. Crossing faults located in the stream valleys are thought to account for these offsets and changes in orientation.

Small-scale structural features in the rocks adjacent to the NWBF may reflect a wide variety of motions during the complex deformation of those rocks. Boone (1981) has recently reviewed the results of a field conference held during the summer of 1980 and attended by G.M. Boone, E.L. Boudette, M. Bronston, S. Serra and this author. Boone's earlier work had led him to the conclusion that the NWBF represents a thrust dipping to the northwest. Boudette's work led him to postulate a thrust from southeast to northwest. My work suggested that the NWBF represents a normal fault, north side down. The conference showed that clear evidence was available on the outcrop scale for all three models, which led Bronston and Serra to propose a fourth model to accommodate all the evidence.

They suggested that a right-lateral strike-slip master fault with an irregular surface would generate synthetic and antithetic shears as well as both normal and thrust surfaces, each accompanied by folds with appropriate rotation sense. If one wants to account for individual features at scattered outcrops along the length of the fault zone, this model is very attractive, but it seems to fail to account for the sharp changes in orientation of segments of the fault and for their offsets. This author suggests that the dominant movement on the NWBF during the deformation of the post-Upper Silurian section may well have been right-lateral, producing many of the disparate small-scale structures, but that the most recent movement (post regional deformation) was normal faulting on that existing fault surface, producing the existing map pattern.

In the Attean quadrangle, Upper Silurian conglomerates contain clasts of the Attean Quartz Monzonite and Chain Lakes lithologies (Albee and Boudette, 1972). This tells us that this core of the Boundary Mountain Anticlinorium was rising relative to the basin to the northwest which was sinking to accept sediments to form the post-Upper Silurian section. During the deformation of the post-Upper Silurian section (the Acadian Orogeny), northwest-southeast compression tightly folded the sedimentary rocks and perhaps thrust them up onto the basement complex to the southeast. Perhaps this contact had a relatively steep angle and a strike-slip component to its motion, but neither a reverse fault or an unconformity would account for offsets in the contact or abrupt changes in orientation of the contact.

The model preferred by this author is one involving post-Acadian uplift of the granitic massif (just like the post-Taconic uplift), with a moderately high-angle normal fault developing along the northwestern border of the massif. Different rates of uplift along the length of the NWBF would produce tear or hinge faults, allowing segments of the fault to rise at faster or slower rates than neighboring segments. Segments of the NWBF which rose faster would be offset to the northwest. Segments where one end rose faster than the other end would have a different strike than adjoining segments. This style of faulting could account for the mapped pattern of the NWBF.

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ROAD LOG

Field work for this trip was done during a four week period in the summer of 1978 while active woods operations were underway. Since that time, timber cutting has ceased and the access road to the region has not been maintained. In June of 1983 the area was revisited for the purpose of preparing this trip log and the author was startled by the extent to which spring rains had damaged the roads. Bridges were washed out and erosion gullies across the road were numerous. It is hoped that by the time of this scheduled trip that the road will be in significantly better condition. Particularly bad washouts are parenthetically noted in the trip log.

The assembly point is in the northern part of the town of Jackman on the west side of Route 201 at the parking lot of Dolly's Food Store (IGA) and the Jackman Post Office.

Mileage

0.0	Turn left onto Route 201 heading north.
4.4	Turn left on a paper company haul road known as both the Crocker Pond Road and the Holeb Road.
6.0	Cross Brady Brook.
8.1	(Careful - washout.)
9.5	Cross West Branch Sandy Stream. (If the bridge is still out, cross the shallows on the downstream side.)
11.2	Cross stream.
11.7	(Careful - washout.)
12.0	(Careful - washout.)
12.4	(Careful - washout.)
14.4	(Careful - washout.)
15.0	Mud Pond is on the south side.
16.5	Junction with the road to Turner Pond. Turn right (north).
17.8	Excellent spring on the right side.
18.1	Bear left at the fork.
18.3	Park along the side of the road.

Stop 1. Exposures of the Attean Quartz Monzonite occur in and to the west of the road. The rocks are porphoritic with sub- to euhedral potassium feldspar phenocrysts generally 1 cm. in diameter. The alteration of biotite to chlorite has produced a weak north-easterly trending schistocity. Numerous nearly vertical mineralized

fractures have prominent orientations near N2E and N44W. Two similarly mineralized shear zones trend N40E and N80W. These rocks are the least deformed example of the Attean Quartz Monzonite which will be seen on the trip, and they are also furthest from the Northwest Boundary Fault (1 mile).

Continue driving north along the road.

19.0 Continue past the entry road on the right which goes into the west end of Turner Pond.

19.35 Park along the road. There is an old grown-over woods road on the north side which starts into the woods on a bearing S80W. This road bends all around, roughly following a brook northwesterl for 305 paces (1600 feet). At that point, an old road forks off S75W to the left. Take that fork for 142 paces (750 feet) where a well established game trail bears off N60W. Follow the game trail roughly parallel to the slope for 54 paces (285 feet). Continue to contour parallel the slope for 108 paces (570 feet) heading generally N70W. Then head N60W for 30 paces (155 feet) dropping from open maple forest down into a flat. Now follow the contour S65W for 70 paces (370 feet). This brings you to the east end of a gully system made by the Northwest Boundary Fault. (If you should get lost, simply head WNW uphill. If the rocks are sheared Attean Quartz Monzonite, keep bending north. If they are brown weathering limestone, you are in the fault zone. If they are black phyllites, you've gone too far north so bend southerly.)

Stop 2. Three principle lithologies are exposed along the linear topographic depression which trends S54W over the crest of the hil for a distance of just under 2,000 feet. Ths southeast side of the depression is flanked by strongly mylonitized Attean Quartz Monzonite. Within the depression are numerous outcrops of limestone with deformation varying from strong to unobservable. The northwestern side of the depression is made up of slaty and phyllit rocks of the Seboomook Formation. Figure 2 shows the distribution of mapped outcrops but represents only a modest percentage of the total number of exposures.

The sheared rocks of the Attean Quartz Monzonite have a strong fracture foliation which typically trends more northerly (N25-40E) than does the northwestern contact of this unit. This relationshi is persistent for 10 miles along strike to the southwest, independent of whether the rocks are part of the Attean Quartz Monzonite or part of the Chain Lake massif further to the southwest (see Figure 3). The easternmost outcrop shown in Figure 2 is the first outcrop to be examined. Intensely sheared Attean Quartz Monzonite exhibits fractures trending N33E,74W which have chattery surfaces indicating normal movement. Shear fractures measured at the east end of the outcrop trend N40E,55W. The intersection of these less steeply dipping fractures may produce the chattery character of the more steeply dipping fractures, negating the possibility of interpretation of the sense of motion. A point for discussion.

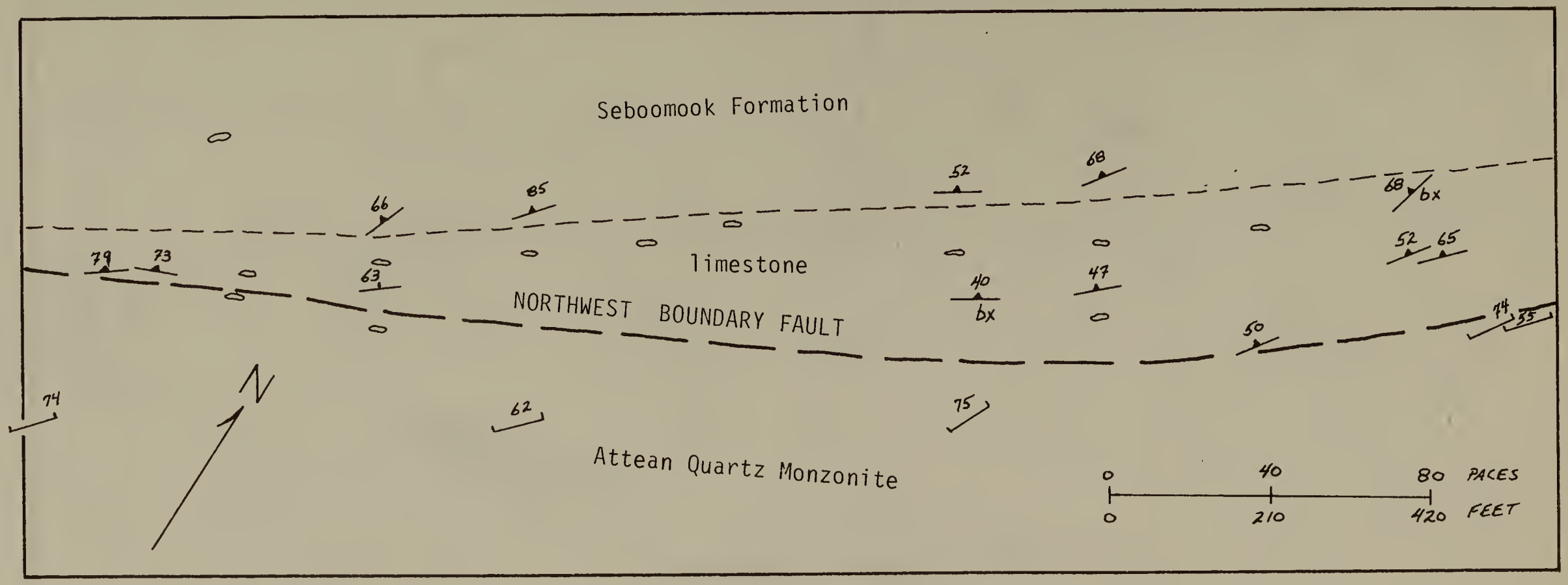


Figure 2. Detailed map of a sliver of Upper Silurian limestone located on the crest of the hill east of East Branch Gulf Stream in the Skinner NE 7½-minute quadrangle

CLEAVAGE AND FRACTURE FOLIATION MAP

Northern Boundary of the Boundary Mountain Anticlinorium



EXPLANATION

- Ds — DEVONIAN SLATES
- Sl — SILURIAN LIMESTONE
- fv — FELSIC METAVOLCANICS
- mv — MAFIC METAVOLCANICS

FAULTS

LITHOLOGIC CONTACT

FRACTURE FOLIATION

CLEAVAGE

SCALE 1:24000

...ARY POND QUADRANGLE
...NNER QUADRANGLE

N

Figure 3A

96

Figure 3B

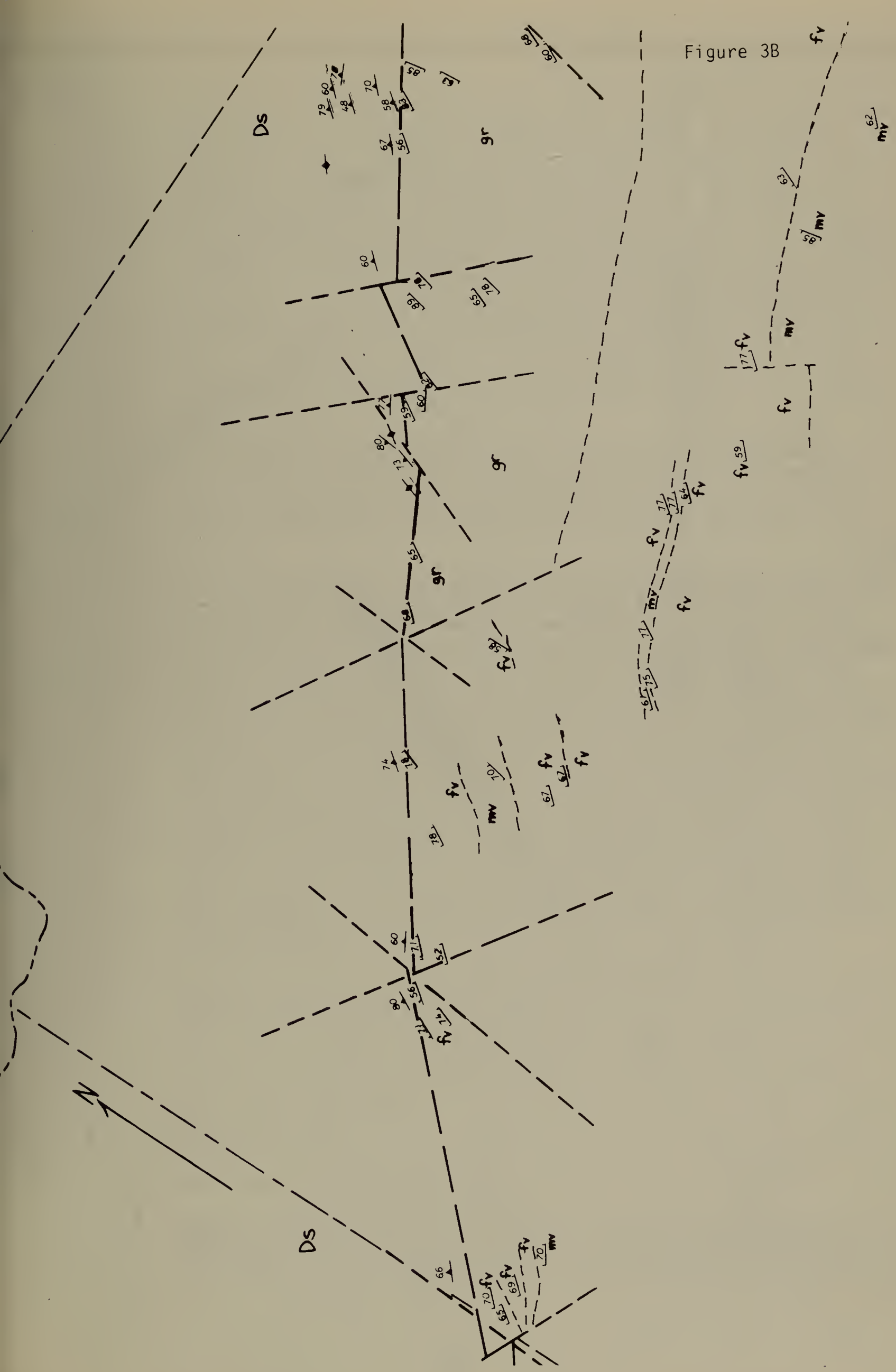
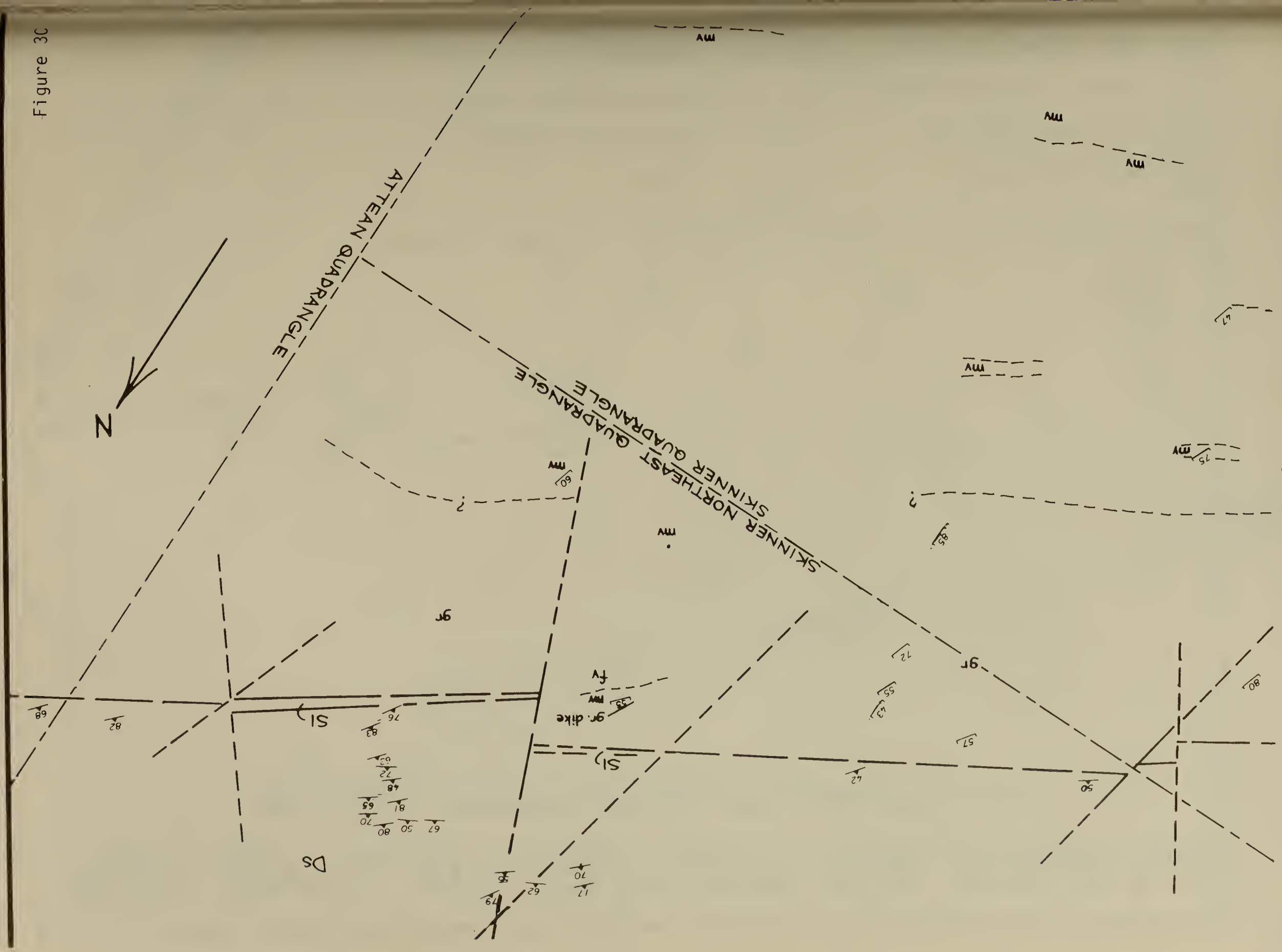


Figure 3C


ATTEAN QUADRANGLE
SKINNER NORTHEAST QUADRANGLE
SKINNER QUADRANGLE
N
mv
gr
gr. dike
fv
Sl
Ds

A traverse heading S60W down the contact zone with side traverses periodically will provide an opportunity to see the detailed lithologic and structural characteristics of the exposures. The number of outcrops visited will be determined by the size and energy of the group. Particular interest is in the origin of the brecciated rocks within the limestone belt. The following have been observed: 1) clasts of mylonitized Attean in a strongly foliated limestone matrix, 2) clasts of limestone in mylonitized Attean, 3) angular limestone fragments in a micaceous matrix, and 4) angular to sub-rounded clasts of Chain Lakes granofels (PЄ), mylonitized Attean (O) and limestone (S) along with calc-silicate rocks and various granitic rocks, all in a strongly foliated quartzo-feldspathic matrix. It is this final breccia which constitutes the westernmost end of the outcrop belt and is the principle gathering point for this stop.

The polymictic breccia described as 4) above is easily located as a major cliff several meters high and 60 meters long, dropping down to the southeast. A cross-sectional view at the west end of the outcrop suggests a preferred orientation of N52E,79N for the clasts.

If the Attean Quartz Monzonite and Chain Lake massif were exposed during the Silurian and yielded pebbles and cobbles to form a conglomerate, then the limestone fragments in this breccia must be accounted for. Deposition followed by uplift of some limestone could have occurred, providing a source for these clasts. Presumably, the depositional surface would have been close to horizontal or dipping gently to the north (away from the source), but it now dips steeply. An unconformity of conglomerate on crystalline basement could be folded into a steeply dipping orientation, but 100 meters to the south the Attean Quartz Monzonite shows no evidence of deformation whereas 50 meters to the south the rocks of this unit are chloritized and well foliated (N40E,74W). Off to the north, bedding in the Seboomook dips 25NW, and locally bedding in the limestone dips 30-50NW. It is the author's interpretation that the breccia here is both a depositional (correlative with the Foxes Camp conglomerate of Albee and Boudette, 1972) and a fault breccia (representing the normal movement (NW down) on the Northwest Boundary Fault).

Return to the vehicles by following the fault NE and picking up the traverse which brought you in.

Continue in the same direction (northeast) on the gravel road.

20.3 Bear left at the fork.

20.9 Cross the brook.

21.0 Park along the side of the road.

Stop 3. Outcrops along the roadside (east side) are sheared Attean Quartz Monzonite which is strongly chloritized giving the rock a pale green color on fresh surfaces. This chlorite and recrystallized

sheared feldspars produce a weak foliation oriented N58E,65N. Three pervasive brittle fracture sets can be seen. The early two are oriented N17E,55W and N63E,60N, and the last set to develop is oriented N42E,72N. Finally, one slickensided fracture cleavage surface has been seen here oriented N37E,41N with stepped chlorite and quartz showing normal movement (N side down).

Continue N on the gravel road.

21.3 Park along the roadside to examine outcrops in and on the east side of the road.

Stop 4. Outcrops of Seboomook Formation show an extremely complex pattern of smale-scale deformation features. Early cleavage is parallel to lithologic layers which may or may not represent bedding. (Elsewhere in this formation false bedding is common as a result of transposition of finer-grained material along steeply dipping cleavage.) The trend of this early cleavage is N48E,80N, quite compatible with the regional cleavage patterns in this formation. Folds typically have amplitudes nearly equal wavelengths, and only rarely can a sense of "handedness" be observed. Fold axes have a variety of orientations; measurements include 44,N75W for a right-handed, chevron-type fold, 30,N45W for open antiforms and synforms, and 16,N33E for a set of waves which are truncated by a N37E,75N fault surface. The continuation of this fold set has not been located. Our fold set has steeply plunging axes (83,N3E) and shows clear chevron drag structure with a right-lateral sense of motion (long limbs N48E,78N). It can not be shown that these folds are related to the most recent movement on the Northwest Boundary Fault.

Turn around and proceed south.

21.45 Optional stop. Park along the road and walk N50E for 340 paces (1800 feet) to the crest of the hill.

Stop 5. Outcrops to the northwest are Seboomook Formation with well-developed slaty cleavage trending N33-63E, 45-65NW. A second cleavage can be seen oriented N18E,79W as axial planes to small open folds whose axes trend 71,N0E. Outcrops on the southeast side are mylonitized Attean Quartz Monzonite with well-developed fracture cleavage N25-47E, 54-70NW. The contact between these two units is tightly confined (see Figure 4) and strikes N50E based on detailed mapping along strike. The fracture cleavage in rocks near the contact commonly strikes more northerly than does the contact. This same relationship was seen at the second stop and remains poorly understood. Veins of quartz are more abundant in outcrops of both formations near the contact than they are further away.

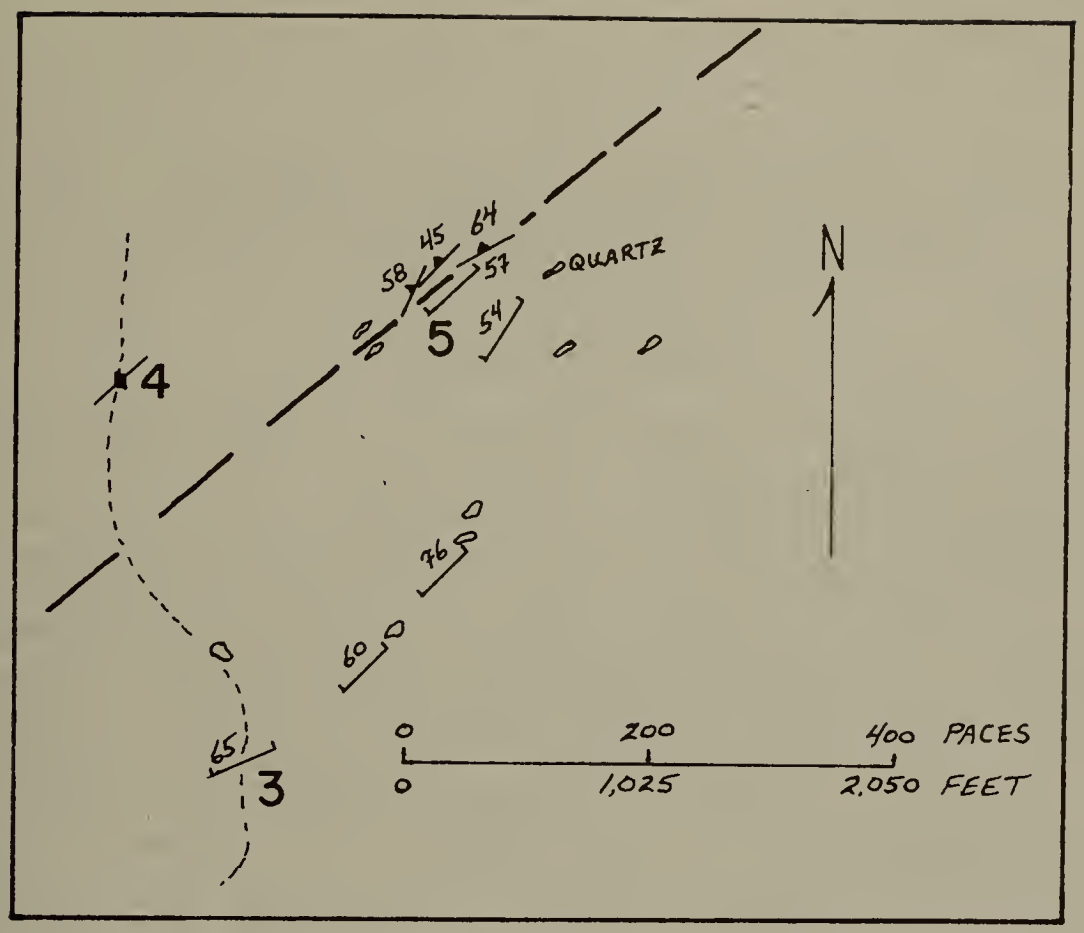


Figure 4. Detailed map of the localities of Stops 3, 4 and 5 northwest of Little Turner Pond in the Attean 15-minute quadrangle

Return to vehicles and retrace your route back to the Holeb Road.

26.1 Junction with the Holeb Road. Turn left and return to Route 201.

38.2 Junction with Route 201. (Excellent roadcuts can be seen on Route 201 to the north if time permits.)

End of trip.

Trip C-1

Petrology of the Central Portion of the Moxie Pluton

By

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General Statement

For the past several years, our group at the Department of Geology and Geophysics, Boston College has been involved in the studies of various aspects of the Acadian igneous rocks located along the northwestern margin of the Appalachian orogenic system in northern and north-central Maine. In this region, 350-420 m.y. before present, a particularly widespread igneous activity lead to the formation of a wide and contrasting assemblage of plutonic and volcanic rocks, ranging in composition from ultramafics to troctolites, norites, diorites, granodiorites and granites and from mildly alkalic basalts to andesites and rhyolites. We envision that integration of the results of the major element geochemistry, trace element geochemistry and mineral equilibria studies will provide further insights into the dynamics of the tectonic evolution of the Appalachians of Northern Maine during the Acadian orogeny.

In this guidebook, we will report chiefly on the field relationships and distribution of the individual petrographic types in a mafic plutonic body. Emphasis is placed on the mineralogical variation and the structural and textural observations within the central portion of the Moxie Pluton. For reasons of simplicity and the keeping this description within the general scope of the NEIGC guidebooks, we will refrain from reporting any geochemical data, which will be published elsewhere. Nevertheless, where it will be appropriate or helpful, we will draw from these results.

Introduction

The Moxie Pluton is a conspicuously elongate body of mafic rocks 75 km long by 3-10 km wide, located near Moosehead Lake in north-central Maine. Trending southwest-northeast, the pluton extends from Moxie Mt across Moxie Pond and Moosehead Lake to its northeast termination in contact with the Katahdin Granite, covering an area of approximately 375 sq km. The age of the pluton has been determined from gabbroic rocks from the central portion of the pluton, dated at 393 m.y. (Early Devonian) by K-Ar biotite methods (Faul, et al., 1963). The pluton intrudes sandstones and slates of the Early Devonian Seboomook Fm., with a prominent contact aureole of biotite-to-cordierite grade hornfels up to 3 km wide.

As shown in Fig. 1, the distinctive outline of the Moxie Pluton defines a series of bulb-shaped centers of intrusion and neck-like zones.

Rocks in the Moxie Pluton range in composition from dunites

and peridotites in the southern bulb through troctolites, norites, and gabbros around Moosehead Lake in the central bulb to quartz diorites and granodiorites in the northeastern neck. Estimates of pressure for the Katahdin and Lexington Batholiths (Hon, 1980), coupled with estimates based on metamorphic isograds published in the Preliminary Geologic Map of Maine (1967), indicate a southwest-northwest post-orogenic regional tilt of about five degrees. Thus, rocks have been uplifted to a greater degree in the southwest, and present exposures represent progressively deeper sections within the crust from northeast to southwest. Therefore, the depth of emplacement for rocks in the southern bulb is approximately 6.5 km greater than for rocks 75 km to the northeast, near the contact with the Katahdin Batholith.

Hon and Roy (1981) presented a regional synthesis for the Acadian Orogeny in northern Maine that accounted for both the occurrence of widespread Acadian plutonic activity in north-central Maine and the changing locus of deepwater sedimentation during Siluro-Devonian time from the Merrimack Synclinorium southeast of the Greenville Plutonic Belt to the Connecticut Valley-Gaspe' Synclinorium northwest of the plutonic belt. As the axis of maximum sedimentation shifted from southeast to northwest, the sense of direction of sedimentary transport concurrently changed from eastward to westward. Located at the southeast margin of Early Devonian graywacke-slates of the Connecticut Valley-Gaspe' Synclinorium, the igneous rocks of the Piscataquis Volcanic Belt (Rankin, 1968) and Greenville Plutonic Belt (Hon, 1980) mark the northwest limit of Devonian magmatism in Maine. Hon and Roy interpret this as evidence indicating that the Greenville Plutonic Belt represent the North American continental margin during the Acadian.

Background

Rocks in the central portion of the Moxie Pluton range in composition from troctolites and anorthosites to norites and gabbros. In hand sample the rocks are light to dark gray, typically medium-grained, and relatively fresh. The dominant phases are plagioclase, olivine, and orthopyroxene, with less but significant clinopyroxene. Hornblende, biotite, apatite, and opaque minerals are minor but widespread. The rocks have a heterogeneous appearance in outcrop, locally displaying cumulate textures that include phase or rhythmic layering, igneous lamination, prominent oikocrysts, and irregular segregations of dominantly feldspathic or mafic material.

Modal and geochemical data support the classification of these rocks into two groups: magnesium-rich and iron-rich (Espenshade and Boudette, 1967). The magnesium-rich rocks, referred to as Mg-rocks in this study, contain olivine or orthopyroxene with optically determined Fo or En contents greater than 50. They are found principally in the western and southern sections of the central portion of the pluton. West of Moosehead Lake, Mg-rocks form the center of the neck of the pluton, extending southwest to the southern bulb at Moxie Mountain. East of the Lake, Mg-rocks outcrop at Lower and Upper Wilson Ponds, forming the southern

"lobe" of the central bulb. Fe-rocks (with optically determined Fo or En contents less than 50) are found along the contacts of the neck of the pluton west of the Lake, encircling the Mg-rocks. East of the lake, Fe-rocks predominate, outcropping in all areas except Lower and Upper Wilson Ponds. Because each of the two rock types has distinctive textural, mineralogical, and geochemical features, they will be discussed separately within each category.

A small siliceous stock, the Beaver Cover Granite, is found in a restricted area around and northeast of Beaver Cove in Moosehead Lake Quad, together with associated aplites, granite dikes, and cordierite-rich rocks. These rocks will be discussed separately from the mafic rocks in the Moxie Pluton.

Textures

Definitions: The terminology used to describe cumulate rocks was first developed by Wager et al. (1960) from their study of the Skaergaard Intrusion. As defined by them, cumulus crystals are precipitated from crystallizing magma and settle at a surface of the magma chamber to form accumulations of discrete crystals. Intercumulus crystals are formed from magma that is trapped in the pore spaces between cumulus crystals. Orthocumulate rocks consist of both cumulus and intercumulus phases. In these rocks, cumulus crystals are euhedral to subhedral and are commonly zoned. The outer zones of the crystals represent intercumulus material of slightly different composition from the original crystal. Intercumulus phases are anhedral and commonly form poikilitic crystals enclosing the euhedral cumulus crystals. Poikilitic crystals may also be zoned. Adcumulate rocks consist of largely unzoned, anhedral to subhedral cumulus crystals, displaying an interlocking mosaic texture, with few or no poikilitic or other intercumulus crystals. Intercumulus material of the same composition as the cumulus crystals diffuses to them; as the cumulus phases continue to grow, they expell from the crystal mush any intercumulus material that is not needed for their growth. Adcumulate textures are best seen in monomineralic rocks such as anorthosites. In mesocumulate rocks, cumulus phases show some adcumulate growth, but retain a small amount of intercumulus material.

Some of the troctolites and pyroxene-anorthosites in the Moxie Pluton have a spotted appearance due to the presence of large poikilitic pyroxenes ranging from one to a few cm in diameter. Following the usage of Jackson (1961) and Irvine (1974), these are called oikocrysts: poikilitic intercumulus crystals, grossly spherical or spongy in shape, that are much larger in diameter than the euhedral cumulus crystals they enclose.

Cumulate textures are readily identifiable when cumulus and intercumulus material form different phases. As the number of cumulus phases increases, the number of differing intercumulus phases decreases; the intercumulus material is more likely to lead to the formation of zoned or overgrown cumulus crystals. With the reduction in number of new intercumulus phases (particularly, prominent poikilitic crystals), cumulate textures become less apparent.

Olivine-bearing rocks in the Moxie Pluton may display a symplectite texture. Where olivine crystals are in contact with plagioclase, a wormy intergrowth of feldspar and pale green to colorless diopside and/or amphibole rims the olivine. The blebs of diopside are no more than a few microns in diameter; the symplectite rim is typically ¼.1 mm wide. Locally, skeletal orthopyroxene mantles the olivine, adjacent to the symplectite. Rarely, symplectite occurs around opaque minerals. Describing a similar symplectite corona in olivine gabbro of the Duke Island Ultramafic Complex, Irvine (1974) considered the texture as evidence for "an approach to local equilibrium." The presence in the rock of two major phases in initial disequilibrium suggests a possible cumulate origin for the rocks, where cumulate textures are obscured by the multiplicity of cumulate phases.

Textures in Mg-Rocks: Cumulate textures seen in Mg-rocks include a spotted texture caused by the presence of abundant prominent oikocrysts, best seen in troctolites at Stop 5 at the center of the neck of the pluton, just west of Moosehead Lake. The oikocrysts are gray, roughly spherical, from one to five centimeters in diameter, set in a lighter gray to white matrix of feldspar crystals (2 mm long). On weathered surfaces, the spots form raised knobs, locally interconnected or roughly aligned to form irregular, discontinuous lineations. The oikocrysts consist of poikilitic clinopyroxenes enclosing euhedral feldspar and olivine crystals, and form 5-25% of the rock. Despite the presence of up to 25% intercumulus material, the spotted troctolites are not simple orthocumulate rocks. Domains of orthocumulate pyroxene texture alternate with domains of feldspar plus olivine that are more adcumulitic in appearance; each domain is only several mm to a few cm in diameter. One sample of adcumulus troctolite from the northern end of Stop 5 contains only 3% intercumulus material. Oikocrysts are also prominent, but less abundant, in anorthosite at Harfords Point west of Moosehead Lake (Stop 8).

Rhythmic inch-scale layering caused by the alternation of narrow bands of mafic- and feldspar-rich rocks is occassionally well-developed at Lower Wilson Pond. Adcumulate anorthosite layers, with a few grains of olivine, grade into orthocumulte troctolite layers in which intercumulus pyroxenes poikilitically enclose feldspar and olivine. Individual layers are irregular and discontinuous, from .5 m to 2-3 m long, and typically only 5-10 mm wide; but paired bands may be up to 5 cm wide. Similar layering is visible at Stop 14, north of Mud Pond, where a weak igneous lamination, caused by subparallel alignment of feldspar laths, is visible in anorthosite layers. The troctolite layers contain poikilitic hornblende and biotite as well as pyroxene.

Elsewhere, phase layering is less well defined. A lineation is sometimes caused by stringers of mafic crystals. In thin section, small olivines appear to be strung out like beads; the effect is produced in part by adcumulus growth of the olivines, fingering into interstices between feldspr crystals. McCallum et al. (1980) describe troctolites in the Stillwater Complex where "banding is defined by subparallel stringers of rounded olivines"

(p. 62).

Orthocumulate textures are dominant in peridotites found in the southern part of the pluton, on the west shore of Lake Moxie: large poikilitic pyroxenes enclose euhedral cumulus olivines. In contrast, adcumulate textures are dominant in the anorthosites at Stop 8 despite the presence of oikocrysts of pyroxene up to 7x10 cm large. In his discussion of the Stillwater Complex, Hess (1960) describes an anorthosite of 90% plagioclase and 10% poikilitic hypersthene that forms large spots up to 3 cm in diameter.

Textures in Fe-Rocks: The most prominent textures in Fe-rocks are igneous lamination and symplectite growth; phase layering plays a subordinate role. The igneous lamination may be accompanied by a linear orientation of feldspar crystals in the plane of lamination (Visher, 1960). The lamination is best seen in norites and gabbro-norites east of Moosehead Lake where olivine is absent or present only in subordinate amounts: the rounded shape of olivine grains tends to disrupt the lamination. At Stop 13, norites with igneous lamination share characteristics of some adcumulate rocks: only two major phases are present, with minor opaques, and there is little interstitial material. In contrast, at Stop 15 on Lily Bay, gabbro-norite with a weak igneous lamination contains 15% poikilitic biotite and hornblende, indicating the orthocumulate nature of the rock.

An igneous lamination accompanies phase layering in rocks from the southern end of Stop 9 west of Moosehead Lake. In one sample, inch-scale layers of adcumulate gabbro-norite show a strong lamination in which pyroxenes and apatite are also oriented in the plane of lamination. These layers alternate with layers rich in intercumulus biotite and hornblende (10% hydrated phases) and devoid of lamination. There is also a suggestion of sorting: apatite grains are largest and most abundant near the "bottom" contact of the adcumulate layer with the orthocumulate layer. A similar sorting of olivine crystals is observed in phase-layered noritic rocks from outcrops near Beaver Cove. Fine-grained layers 4-5 mm wide alternate with 1-5 mm wide bands of mafic minerals and layers a few cm wide of olivine norite. Grain sizes in the olivine norite layers are somewhat larger than in the other layers, but toward the contact with the norite layer, grain sizes and the proportion of olivine crystals decrease. The contacts between layers are sharp.

Symplectite texture is ubiquitous in olivine-bearing Fe-rocks, increasing in prominence as the proportion of modal olivine increases. It is best seen in Fe-rocks at the northern contact of the neck of the pluton, west of Mooshead Lake, where both orthocumulate (Stop 2) and adcumulate troctolites are found.

To summarize, textures in Mg-rocks include prominent oikocryst textures and rhythmic, inch-scale, phase layering, with minor but significant monomineralic adcumulate textures (massive anorthosites). Fe-rocks may display phase layering, but the dominant texture is an igneous lamination of feldspars that is strongest in adcumulate rocks where other phases, notably pyroxenes, are also aligned in the plane of lamination.

Mineralogy

The minerals in mafic rocks of the Moxie Pluton may be classified as cumulus or intercumulus phases, according to criteria defined above. While olivine and plagioclase occur almost exclusively as cumulus phases, ortho- and clinopyroxene, apatite, and opaque minerals may form either cumulus or intercumulus grains. Hornblende and biotite occur almost exclusively as intercumulus phases; quartz is intercumulus only. Alteration products include fibrous amphibole, serpentine, chlorite, talc, carbonate, sericite, and kaolinite; but except for ultramafic rocks, the rocks are relatively unaltered.

Mineralogy of Mg-Rocks: In Mg-rocks where olivine and plagioclase are the only cumulus phases, as in the spotted troctolites at Stop 5, olivines are commonly mantled with a skeletal rim of orthopyroxene. The troctolites are feldspar-rich, consisting of 60-70% plagioclase (An 70, optically determined) and 10-30% olivine (Fo 70-75). Intercumulus oikocrysts of clinopyroxene make up 5-20% of the rock, with minor intercumulus red-brown biotite (.5-3.0%) and hornblende (.5-5.0%) that may mantle olivine or intercumulus opaques (.5%). Olivine and plagioclase in layered troctolites are similar in composition. Paired anorthosite-troctolite layers may be more feldspathic than the spotted troctolites (70-80% plagioclase), or more hornblende- and biotite-rich.

Clinopyroxene appears as a cumulus phase at Stop 4, where olivine gabbro contains 20% cpx. Euhedral grains of cpx are clearly cumulus in origin: the subophitic texture of other grains of cpx may represent orthocumulate overgrowth or adcumulate extension of original cumulus crystals. The An and Fo contents of the plagioclase and olivine are sismilar to those of the spotted troctolites that lie west and north of Stop 4. In contrast, at Stop 6, feldspathic olivine gabbro contains plagioclase with An 56-60 and olivine with Fo 60-65 as cumulus phases with cpx.

While cumulus cpx occurs in outcrops on both sides of Moosehead Lake, cumulus opx in Mg-rocks occur only in the eastern half of the pluton. Stop 10 at the mouth of Beaver Cover, is a fine-grained hypersthene gabbro with cumulus cpx and opx of En 60. Some cpx are subophitic, suggesting continued intercumulus growth on cumulus crystals. However, cumulate textures are not clearly defined, except in isolated, irregular patches of coarse-grained feldspathic rock up to 2 m long, with En 50-55 and An 50. Subparallel alignment of pyroxenes and some plagioclase crystals is interrupted by poikilitic hornblende and biotite crystals 3-5 mm in diameter. Trace amounts of intercumulus apatite are present; and the sample contains 4% intercumulus quartz. Contacts between the coarser-grained material and the hypersthene gabbro are sharp.

Sample at Stop 12 is a norite with 53% plagioclase, 43% opx of En 65, and minor hornblende, biotite, and opaques. Feldspar and some of the pyroxenes show adcumulate texture. The majority of pyroxene crystals, however, are subophitic to nearly oikocrystic. Large pyroxenes that look like single, sub- to euhedral crystals

in plain light, are revealed as aggregates of small, individual crystals under crossed nicols. This suggests that pyroxene is a cumulus phase that underwent continued growth from intercumulus material. It should be noted that where orthopyroxene is a cumulus phase, olivine is rare or absent. There is no olivine in samples at Stops 10 nor 12.

Mineralogy of Fe-Rocks: Fe-rocks are predominantly norites and gabbro-norites with cumulus plagioclase, orthopyroxene, clinopyroxene, apatite, and opaques. Fine-grained norites at Stop 13 consist of 60-65% plagioclase (An 52-57), 25-30% opx (En 40-50) and 5% opaques. Minor hornblende, biotite, and trace amounts of quartz are present as intercumulus grains. In these norites and gabbro-norites from Stop 15 and from near Tussle Lagoon (Espenshade's outcrop 562, 1972), fibrous amphibole identified as cummingtonite appears as an alteration product, chiefly at the expense of orthopyroxene.

Other olivine-free norites and gabbro-norites east of Moosehead Lake typically contain 55-60% plagioclase (An 55-65), 20-25% orthopyroxene (En 35-50), up to 10% clinopyroxene, opaques, and up to 1.5% apatite as cumulus phases. In addition, rocks from some of the outcrops on the eastern side of Beaver Cove, contain a few euhedral crystals of primary cummingtonite. Interstitial quartz may be present in amounts up to 2%. Intercumulus poikilitic hornblende and biotite are ubiquitous. The gabbro-norites extend from Beaver Cove south to Sandy Bay, east to Prong Pond, and north to Tussle Lagoon and North Brook.

Rocks with abundant olivine occur only west of Moosehead Lake. The olivine rocks have an An content of 55-60 with Fo of 30-35. Symplectite texture is ubiquitous. The rocks are coarser-grained than the norites east of the lake; and En contents are lower (35-40 vs 40-50). Outcrops at Stop 9, located at the southern contact of the pluton, are heterogeneous olivine norites containing 45-65% plagioclase, 5-20% olivine, and 5-25% orthopyroxenes. Up to 2.5% cumulus apatite crystals are common; trace amounts of quartz are present. At the northern contact of the pluton, quartz is absent from iron-rich troctolites, and orthopyroxene occurs only as an intercumulus phase mantling olivine.

Comparison of Mineralogies: Mg-rocks are characterized by the predominance of cumulus plagioclase and olivine with intercumulus clinopyroxenes. Orthopyroxene usually appears only as a skeletal rim on olivine. Hornblende, biotite, and opaques together typically make up 3% or less of the rock. Apatite is absent or present only in trace amounts, and potassium feldspar is nowhere present. Plagioclase, olivine, and orthopyroxene compositions vary within a fairly narrow band: An 65-75, Fo and En 70-75. Lower values are reported only for those rocks in which pyroxene is a cumulus phase (En 60-65). In contrast to the more restricted range of the Mg-rocks, Fe-rocks commonly contain four or five cumulus phases, including opaque minerals and apatite. An, Fo, and En contents are all considerably lower than those in Mg-rocks: typical mineral compositions are plagioclase An 50-60,

orthopyroxene En 35-45, and olivine Fo 30-40. Orthopyroxene is more abundant than clinopyroxene: cpx oikocrysts, common in Mg-rocks, are absent in Fe-rocks. Instead, poikilitic clots of biotite and hornblende make up 5-10% of Fe-rocks. Trace amounts of interstitial quartz are found in rocks without olivine; like Mg-rocks, however, Fe-rocks contain no potassium feldspar.

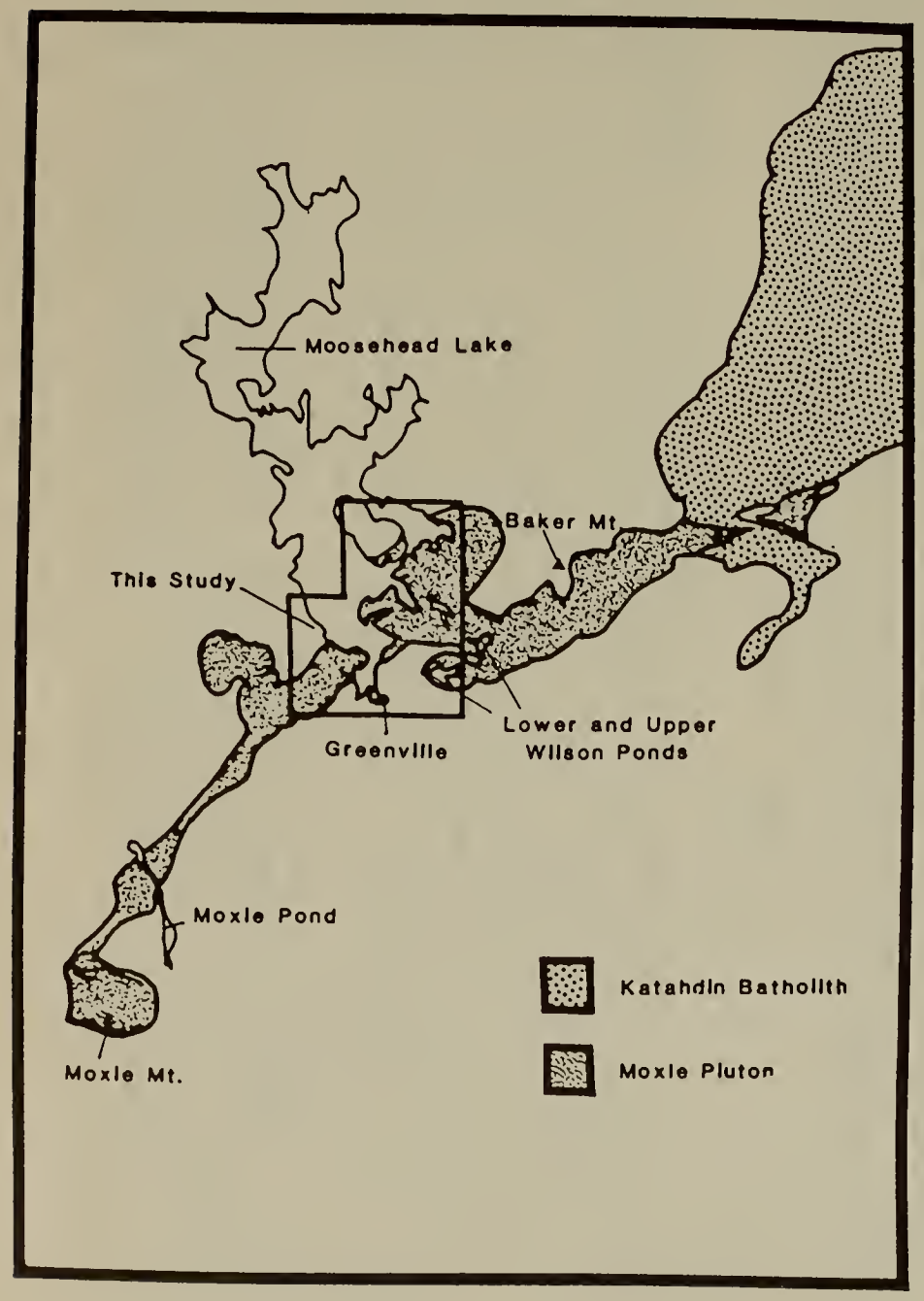
No rocks with orthopyroxene composition of En 50-60 were found in this study. Espenshade (1972), however, reported the occurrence of such pyroxenes in gabbro-norites from east of Moosehead Lake. These rocks share certain characteristics with the Fe-rocks. Around Prong Pond gabbro-norites contain one or more of the following: 5% interstitial hornblende plus biotite, 5-8% opaques, and interstitial quartz. A variable to strong igneous lamination may be present. Along Lily Bay, a laminated norite with abundant apatite and opaques was described by Espenshade (1972).

Granites and Granitic Dikes

Siliceous rocks in the Moxie Pluton include the Beaver Cove Granites with associated aplites and granitic rocks. The granites are best seen northeast of Beaver Cove, where granite outcrops trend north for approximately one kilometer. At Black Point, boulders of brecciated norite in a groundmass of granite indicate the intrusive nature of the Beaver Cove-Moxie Pluton relationship. The granites are a light buff color, weathering to a rusty orange, with a medium- to fine-grained sugary texture. The lack of mafic phases is obvious in outcrop. Stop 13 shows samples which contain 30-35% oligoclase, an equal amount of quartz, and 30% microcline, with minor biotite and alteration products. Some plagioclase grains are rimmed with myrmekite. Scattered clusters of tiny grnet crystals are rare but visible in samples of the aplite, with bleached zones around the clusters. The aplite is similar in composition to the granites, with slightly more microcline and slightly les plagioclase.

Vertical granite dikes that intrude outcrops at Stop 12 show a lineation from alignment of biotite flakes parallel to the walls of the dikes, caused by flow foliation. The dikes are 20-30 cm wide, speckled white to light gray in color, and medium- to fine-grained (grains .5-1.0 mm in diameter). The fine-grained, foliated, quartz-rich material that occurs at Rum Cove on Lower Wilson Pond appears in thin section to be the same kind of granitic dike rocks as samples granitic dikes of Stop 12.

A dark gray, aphanitic dike rock intrudes also the norite cliffs and ledges along the north shore of Prong Pond. The dikes are a few cm wide, with apophyses into the norite. Locally, biotite phenocrysts are visible along the margins of the dikes. In thin section, the rock is fine-grained (grains less than .2 mm diameter), with an equigranular, sugary texture like that of an aplite.


Moosehead Lake
Baker Mt
This Study
Lower and Upper
Wilson Ponds
Greenville
Moxie Pond
Moxie Mt.
Katahdin Batholith
Moxie Pluton

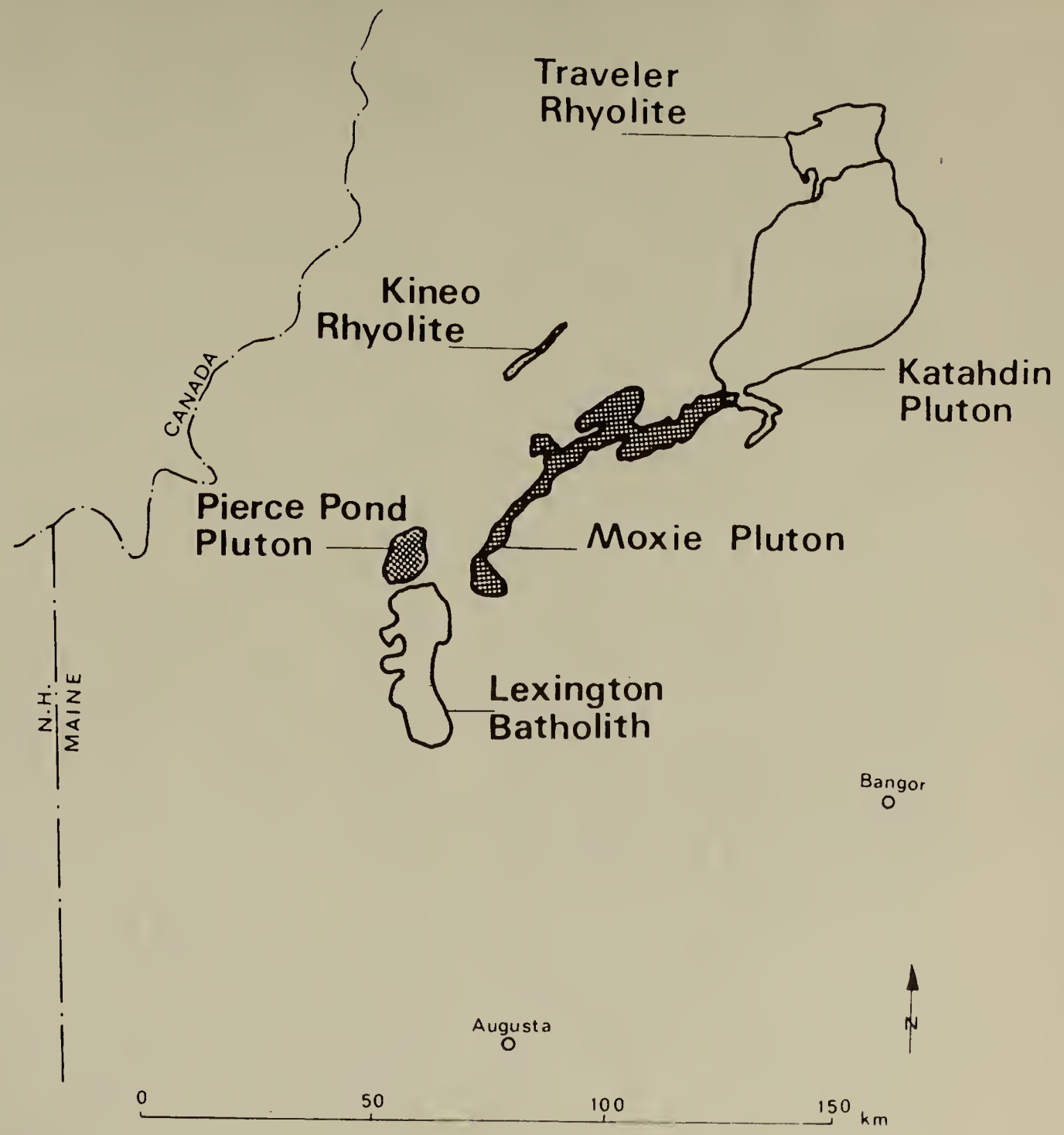

Traveler
Rhyolite
Kineo
Rhyolite
CANADA
Katahdin
Pluton
Pierce Pond
Pluton
Moxie Pluton
N.H.
MAINE
Lexington
Batholith
Bangor
Augusta
N
0
50
100
150
km

Plate I. Central Portion of the Moxie Pluton

● ○ Magnesium-Rich Rocks

■ □ Iron-Rich Rocks

Open Forms, No Geochemical Data

★ Granites

△ Granites Mapped by Espenshade, 1972

◉ Seboomook Fm.

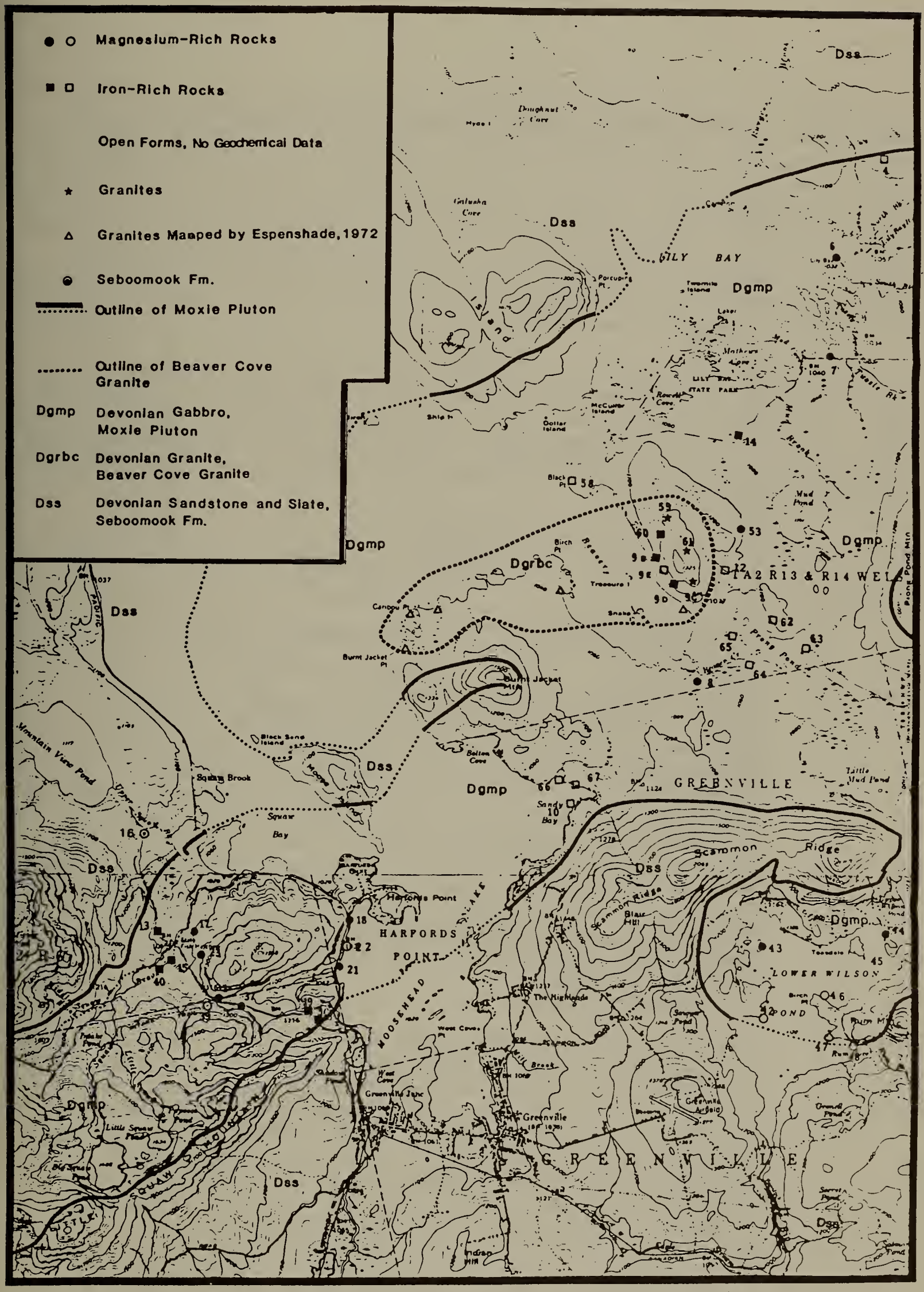
Outline of Moxie Pluton

Outline of Beaver Cove Granite

Dgmp Devonian Gabbro, Moxie Pluton

Dgrbc Devonian Granite, Beaver Cove Granite

Dss Devonian Sandstone and Slate, Seboomook Fm.



SCALE 1:62500

CONTOUR INTERVAL 20 FEET

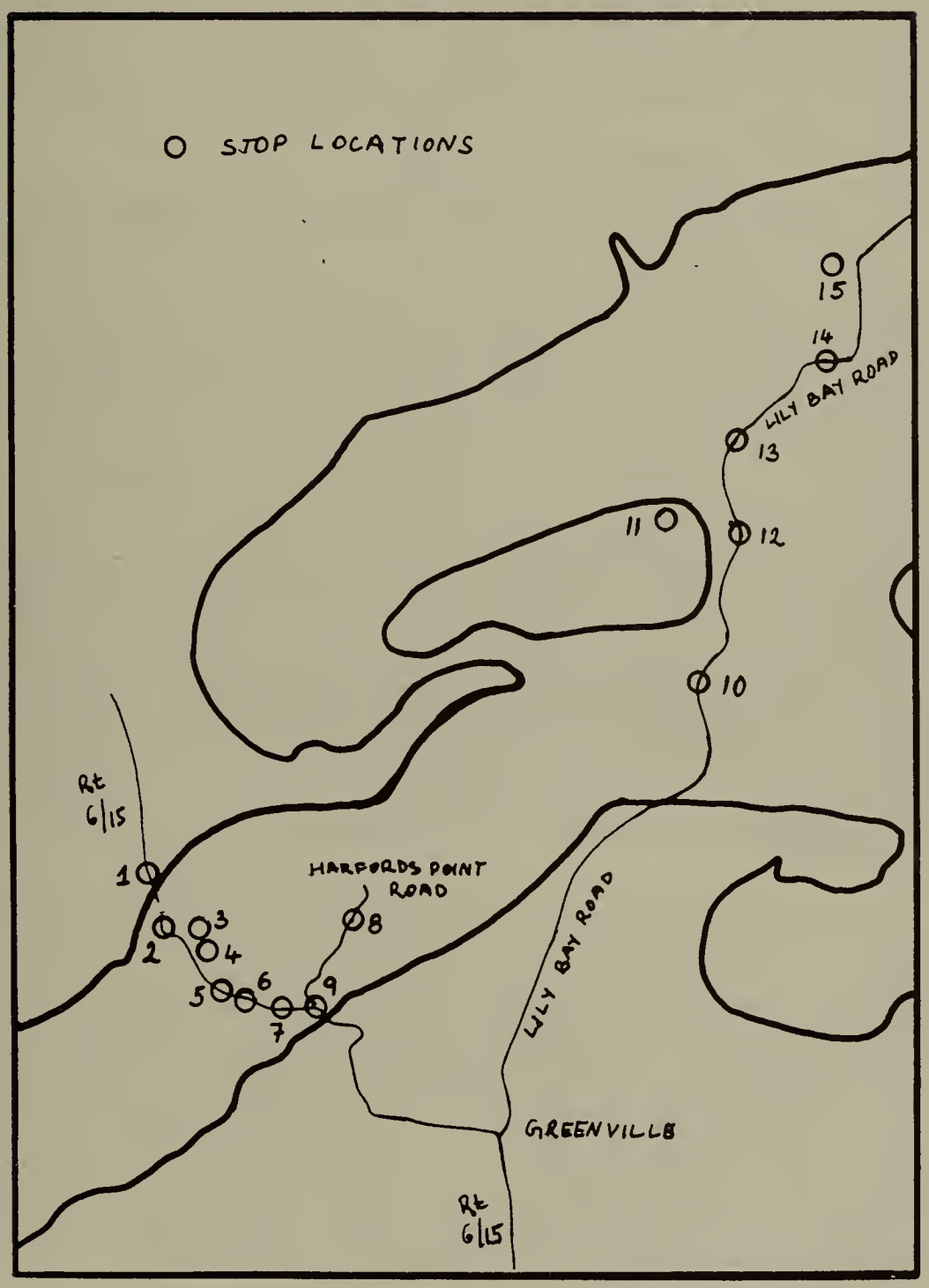
ROAD LOG

Assemble at 8:30 AM in the parking lot of the Squaw Mtn. Ski Lodge at the base of the ski resort area. The lodge is located at the end of the well marked side road, about 6 miles NW from the center of the town of Greenville. If traveling north from Greenville, turn left (West) onto the Squaw Mtn. Ski Resort road and proceed for 2.2 m toward the lodge. Essentially all trip stops will be alongside major highways, either on Rt. 6/15 or Lily Bay Road (Baxter State Park Road). As these roads may be well travelled highways, up to eight vehicles will be available for the trip. Please consolidate into as few vehicles as possible at the beginning of the trip; the total distance travelled will be short and the parked cars can be readily reached at the end of the trip. Please bring your own lunch bags.

This field trip consists of two legs. The first leg will examine outcrops across one of the narrower sections of the Moxie Pluton (about 2.5 mi wide). The second leg will visit outcrops along a SW-NE cross-section through the so-called "Greenville bulge". All localities of this trip are either in the southern portion of the Moosehead Quadrangle or in the nortion portion of the Greenville Quadrangle.

LEG 1

Mileage		
0	0	From the Ski Lodge parking lot go toward Rt. 6/15.
2.2	2.2	Junction with Rt. 6/15. Turn left (south) onto Rt. 6/15.
2.3	0.1	STOP 1: CONTACT METAMORPHOSED COUNTRY ROCKS NEAR THE NW CONTACT WITH THE MOXIE PLUTON. Andalusite to sillimanite grade metamorphism of a well laminated monotonous sequence of rythmically layered, graded beds (graded begging is usually several inches thick). The shaly zones within the graded beds contain in the bedding plane clearly visible, large elongated andalusite porphyroblasts surrounded by the groundmass, consisting of biotite, quartz, feldspars and ocasionally also of sillimanite. Common stringers of pegmatoidal zones are filled with quartz, feldspar, andalusite and biotite. These pegmatoidal zones are believed to have their origin in the dehydration reactions upon heating of the sedimentary rocks. Proceed south on Rt. 6/15.
2.65	0.35	Outcrop on the west side of the road. STOP 2: ORTHOCUMULATE OLIVINE NORITE TO HYPERSTHENE TROCTOLITE OF THE MARGINAL ZONES. This rock is a typical representative of the marginal zone in this area. The rock is medium grained, discretely laminated due to an alignment of mineral grains (N50E, 40°S), contains occasional small fragments of country rock inclusions and infrequent pegmatitic veinlets and nests. Prominent sulfidic eliptical stringers, 2-10 inches long and about 1 inch wide, are easily recognized by their rusty appearance on the weathered surfaces or by a pyrrhotitic yellowish color on the fresh surfaces. The sulfidic stringers are well developed along both sides of the contacts and will also be observed at Stop 9. Mineralogically, this rock is made of approximately


STOP LOCATIONS
15
14
LILY BAY ROAD
13
12
11
10
Rt 6/15
1
2
3
4
5
6
7
8
9
HARFORDS POINT ROAD
LILY BAY ROAD
GREENVILLE
Rt 6/15

50% cumulitic labradorite feldspar, 15-20% hypersthene, 10-20% fayalitic olivine, 5% each of hornblende and biotite, 3% apatite and about 7% ilmenite and other opaques. Biotite and hornblende are present as intercumulus phases forming larger, about nut-sized poikilitic phenocrysts with olivine, opx and plagioclase being the included minerals. Clinopyroxene is virtually absent, while it is abundant within the central portion of the pluton.

Continue southward on Rt. 6/15.

3.15 0.50 Turn left (east) and proceed toward the Bureau of Forestry's Squaw Brook Campsite. Bear left at the upcoming fork.

3.35 0.20 Outcrop about 100 ft. on your left side (away from the stream). STOP 3: MEDIUM-GRAIN GRADED TROCTOLITE. Shallowly dipping sequence (N05E, 25°S-toward the center of the pluton) consists of a series of reptitious motifs, each about 10-20 inches thick. The upper part of each motif is made of a medium to coarse grained troctolite with no, or only few, poikilitic phenocrysts. The grain size may be up to 10 mm large. By contrast, the lower part of the motif consists of fine grained (less than 1 mm) troctolites with abundant, 3-5 cm large, poikilitic phenocrysts (oikocrysts) of clinopyroxene. Mineral constituents are cumulitic labradorite-bytownite plagioclases (60%), forsteritic olivine (20%), and intercumulus clinopyroxene (10-15%) and about 1% each of opaques, biotite, and hornblende. Orthophyroxene represent less than 2% of the rocks.

Turn around and return to the paved road. The next stop is optional.

3.4 0.05 For the next optional stop, continue straight south along the now abandoned old section of Rt. 6/15; otherwise, return to Rt. 6/15.

3.7 0.30 Outcrop (small road cut) on left (east). STOP 4: DISCRETELY LAMINATED OLIVINE GRABBRO TO TROCTOLITE. Discrete lamination and discontinuous layering is caused by crystal alignment, grain-size variation, and by alternating plagioclase-rich or mafic-rich layers. Medium-grained labradorite-bytownite laths constitute 60% of the rock, forsteritic olivine 5-15%, clinopyroxene 10-20%, opx less than 5%, hornblende about 5%, biotite about 2%, and opaques less than 1%. Lamination: N30W, 20°S.

Make "U" turn and return to Rt. 6/15.

4.10 0.40 You're back on Rt 6/15. Turn left and proceed southward.

4.70 0.60 Road custs along both sides of the road at the highest point on the road. Park cars carefully so that you maintain visibility with respect to traffic coming from both directions. STOP 5: OIKOCRYSTIC-SPOTTED TROCTOLITES. Graded and rythmically well-laminated, leopard-skin like sections can be seen on the east side of the road. Here, 40-80 cm thick "layers" show three characteristic zones. The lowest and the most dominant, Zone A, is typically 30-50 cm thick and consists of fine to medium grained, spotted oikocrystic troctolites with rather abundant oikocrysts. The bulk of the rock is made of 60% labradorite-bytownite plagioclase, up to 20% forsteritic olivine, and 2% or less of opx, hornblende, biotite, and opaques. The average grain size is between .5 and 1 mm. The poikilitic, phenocrystic spots

are up to 2 cm large clinopyroxenes, enclosing largely plagioclase and olivine grains. Modal percentage of cpx is about 15%. Zone B usually 10-25 cm thick is a medium to coarse grained rock of the same mineralogy but with less abundant and larger (up to 10 cm) oikocrystic clinopyroxenes. Zone C, anorthositic troctolite, from 0 to 5 cm thick, contains no oikocrysts and consists of up to 1.5 cm large plagioclase grains (70%), and 20-25% of forsteritic olivine, and 5% of minor cpx, opx and hornblende. The lamination (N70W; 25°N) dips toward the north in contrast to Stops 2, 3 and 4 where the lamination was dipping toward the south. This suggests that the reversal of the dip direction indicates that the center of the pluton may lie between Stops 4 and 5. Oikocrysts generally weather less rapidly than the anorthositic troctolite groundmass. On weathered surfaces the rock takes on at first a mottled microorbicular appearance, leading to a formation of "raised knobs" and eventually resulting in a "free suspension" of the oikocrystics in the sapprolitic matrix.

Continue south on Rt. 6/15.

5.25 0.55 Road cut on the left (east side of the road). STOP 6: (OPTIONAL) MEDIUM TO COARSE GRAINED MASSIVE, SLIGHTLY LAMINATED OLIVINE GABBRO. A cumulitic rock, consisting of semi-parallel laths of plagioclases and equidimensional grains of olivine and larger (1-3 mm) irregular grains of intercumulus clinopyroxene. Orthopyroxenes and opaques are nearly absent; this rock resembles rocks of STOP 3, except that clinopyroxene is somewhat more abundant and olivine is less abundant. Also intercumulus clinopyroxene forms irregular grains rather than oikocrysts. Lamination: N80W, 25°N).

Proceed south on Rt. 6/15.

5.50 0.25 Road cut on the right (west side of the road). STOP 7: COARSER GRAINED, WEAKLY OIKOCRYSTIC OLIVINE GABBRO. This rock is faintly laminated due to an alignment of plagioclase laths (An60) which make up about 60% of the rock. Other minerals include 15-20% olivine (Fo75) and 20-25% clinopyroxene with minor opx and opaques. This tock is transitional between the troctolite of the central zone and the gabbronorites of the marginal zones. Small oikocrystic clinopyroxenes (up to 2 cm large) are here side by side with sulfidic stringers typical of the marginal zones.

Continue south on Rt. 6/15.

5.80 0.30 Turn left east onto a road leading to Harford Points (leaving Rt. 6/15).

7.00 1.20 Rairoad crossing of Canadian Pacific RR.

7.20 0.20 Road is divided by a center island which splits this road into two one-way directions.

7.30 0.10 Proceed to the end of the island and turn around (U turn).

7.40 0.10 Stop just before the hill top on the road. Park cars at the extreme right of the side of the road. Proceed on foot to your right toward the RR tracks. STOP 8: ORTHOCUMULATE OIKOCRYSTIC ANORTHOSITIC TROCTOLITES. Yet another type petrographic type of the central zone of the pluton. The rocks consist of largely fine-grained, anorthositic matrix, with variable amounts of olivine (less than 15%) and variable frequency and size of

pyroxene oikocrysts. Labradorite plagioclase can form up to 80% of the rock, giving it a sugary-white appearance. Pyroxene oikocrysts (10-30%) vary in size from a few mm up to 4 cm. Small amounts of up to a few % of intercumulus biotite and hornblende are often observed.

Return to cars and proceed toward Rt. 6/15.

8.80	1.40	You're back on Rt. 6/15; proceed south (left).
8.85/ 8.95	0.05/ 0.10	Road cut along the western side of the road (left). STOP 9: OLIVINE GABBRONORITE-ORTHOCUMULATE HYPERSTHENE TROCTOLITE. The rock types of this outcrop are essentially identical to the rocks of Stop 2. Both outcrops are very near the contact with the country rock and represent Fe-rocks of the marginal zones. Plagioclase (An 55-65) about 50%, olivine (Fo 30-40) 5-20%, orthopyroxene 15-20%, clinopyroxene few % to 15%. Biotite, hornblende and opaques are present each 2-5%, and an abundant quantity (up to 2%) of apatite.

Proceed toward Greenville.

9.10	0.15	Unexposed contact between the pluton and the country rocks.
9.20/ 9.40	0.10/ 0.30	Country rocks exposing sillimanite to andalusite grade, occasionally highly deformed sequences of graywacke-shale beds, and calcareous siltstones.
12.00	2.60	Town of Greenville, junction of Rt. 6./15 and Lily Bay Road.

LEG 2

0	0	Town of Greenville, junction of Rt. 6/15 and Lily Bay Road. Proceed NE on Lily Bay Road.
1.05	1.05	Scammon Road junction on your right.
5.50	4.45	Road cuts and discontinuous ledges on the right side (south) of Lily Bay Rd. STOP 10: HYPERSTHENE GABBRO, DIORITE AND PLAGIOCLASE-RICH GRANITE. The basic rock type of this outcrop is a fine grained hypersthene gabbro, which appears to be non-cumulitic. It consists of 15% opx, 20% cpx, 55% labradorite, and up to 5% of each of hornblende, biotite and opaques. Noted differences between this rock and the rocks of Leg 1 is the absence of cumulitic texture and the absence of olivine. Other rock types in this outcrop include plagioclase-rich granitic rock (30% quartz, 55% plagioclase-An30, 10% alkali feldspar and 4% biotite). This rock is not related to granites of the Beaver Cove type (Stop 11), but rather, they seem to be a late stage fractionation product of the mafic Moxie magma. Hypersthene quartz diorite, also found in this outcrop, probably represents an intermediate fractionation product between the gabbro and the plagioclase granites. Ledges on the south side of the road, but closer to Greenville, crop oikocrystic norites and coarser diorites.

Return to your cars and continue toward NE.

6.20	0.70	Beaver Cove Marina -- check your odometers.
6.80	0.60	Make a left turn onto Black Point Rd. toward Beaver Cove Camps.
6.90	0.10	Leave Black Point Rd. and turn right onto North Ridge Road. For the

next 1 mile there are infrequent ledges of Beaver Cove granites along the road.

7.80 0.90 Lookout Bluff Road on your right -- continue on North Ridge Road.

7.90 0.10 Ledges along the right side of the road (east). STOP 11: GRANITES AND APLITES OF BEAVER COVE TYPE. Fine to medium grained, buff colored leucocratic granite, and very fine grained pink aplitic dikes are seen here intruding ferrogabbros and ferronorites. Geochemically, these granites and aplites are very different from the Moxie pluton mafic rocks and are genetically unrelated to them. The most likely origin of the acidic magma is by a partial fusion of the country rocks surrounding the pluton. The granites consist of 30% quartz, 30-35% alkali feldspar, 25-35% sodic oligoclase, and up to 5% biotite. Garnets are frequently observed in the finer grained specimens.

Continue driving in the same direction.

8.00 0.10 Turn left (SW) onto Cross Ridge Road.

8.20 0.20 End of Cross Ridge Rd.; turn left (south) on Black Point Rd. and proceed toward Lily Bay Rd.

9.50 1.30 You have completed a loop over the ridge NE of Beaver Cove. You're now back on Lily Bay Rd. Turn left (NE) onto Lily Bay Rd. and proceed toward Lily Bay.

10.50 1.00 Ledges on the left (west side of the road). STOP 12: OLIVINE GABBRO TO NORITES. As we proceeded from Greenville to the northeast, there seems to be a decreasing presence of olivine. This rock outcrops consist of medium to coarse grained olivine gabbro and fine grained norites. Labradorite makes up to 55% of the rock and olivine is less than 10%. Opx in the norites amounts to 40% of the rock with no cpx, whereas in the gabbro the cpx is about 20% and opx only a few %. Biotite, hornblende and opaques can represent up to 5% each.

Continue in the same direction (toward NE).

11.70 1.20 On your left is the entrance to Lily Bay State Park. Continue on Lily Bay Rd.

11.85 0.15 STOP 13: FINE GRAINED FERRONORITE TO FERRODIORITES. This rock is similar to gabbros of the marginal zones north of Greenville, except of a noted absence of olivine and a larger presence of plagioclase. Proceed in the same direction.

15.35 1.50 Outcrops on both sides of the road. STOP 14: OIKOCRYSTIC HYPERSTHENE TROCTOLITE TO OLIVINE GABBRO. This rock is one of the few well developed cumulitic rocks within the Greenville Bulge. It resembles some of the troctolites of the central zone north of Greenville. These rocks consist of 10-15% olivine, 55-65% plagioclase (An65), about 10% of each of cpx and opx, and up to 5% each of biotite, hornblende and opaques. Continue in the same direction.

14.40 1.05 You're now crossing over the South Brook Bridge.

14.70 0.30 Side road on your left (NW). Turn left and proceed to the end of the road at the tip of a small peninsula projecting into Lily Bay of Moosehead Lake.

15.15 0.45 End of road. Ledges along the shores of the lake. STOP 15: FELDSPARTHIC NORITE TO HYPERSTHENE DIORITE. Rocks exposed at this outcrop are transitional between gabbro and diorite. They contain 60-65% plagioclase (An50-55), 15% opx, 5% cpx, 5-10% hornblende, 5% biotite, about 5% opaques, and 1-3% quartz. Significant is the ubiquitous presence of quartz and the increasing amount of biotite and hornblende at the expense of pyroxenes.

Make a "U" turn and return to Lily Bay Road.

15.60 0.45 You're on Lily Bay Rd. -- if you go to Greenville (about 15 miles), turn right (SW); if you go to Millinocket, turn left (NE).

END OF TRIP!

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Trip C-2

STRATIGRAPHY, METAMORPHISM AND GEOMORPHOLOGY OF THE GREENVILLE-ROCKWOOD AREA, MAINE

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INTRODUCTION

In the Greenville-Rockwood area, rocks from Cambrian through Middle Devonian age are exposed within the Moose River Synclinorium, the Lobster Mountain Anticlinorium, and the Roach River Synclinorium. The purpose of this field trip is to show how both original lithology and contact metamorphism influence the topography of this region of Maine.

The Moxie pluton was mapped by Espenshade (1982) and Hon (this volume) has studied its petrology. Boone (this volume) has mapped pre-Silurian rocks in the Lobster Mountain Anticlinorium. The rocks within the Moose River Synclinorium were mapped by Boucot (1961) and Boucot and Heath (1969). We first became aware of the origin of the topography in the Greenville area while mapping the surficial geology for the Maine Geological Survey (Caldwell and Hanson 1975). More recently, students at the Maine Geology Field Camp (Boston University) have mapped these rocks under our direction, furthering our interest in the subject.

STRATIGRAPHY

Detailed descriptions of the rocks within this area are found in the introduction to this volume in Moench and others (1982), and in Boucot and Heath (1969). The rocks within the study area are listed in stratigraphic order and described briefly within Table 1. A simplified geologic map of the area is shown in Figure 1.

Geological Control of Topography within the Structural Belts in the Greenville-Rockwood area

The linear ridges and valleys in the area are the result of differential erosion of Acadian fold belts. Irregularly shaped basins and mountains are associated with plutons and surrounding granofels. The rocks within the cores of synclines are the youngest and least deformed and are also the most resistant (for example, Kineo, Tarratine). The pre-Silurian rocks within the core of the Lobster Mountain anticlinorium are highly deformed, showing more than one cleavage and are relatively non-resistant to erosion.

Moose River Synclinorium

The Kineo member forms a line of steep sided hills and ridges stretching from Brassua Lake to near the northern end of Moosehead Lake. This trend includes the spectacular Kineo Mountain. The toughness and durability of this rock is indicated by the fact that the Kineo (locally called the Kineo flint) was prized by the Indians as a material for arrowheads.

The Tarratine forms a nearly 40-mile-long ridge from Spencer Lake on the southwest, to the northeast end of Moosehead Lake. This ridge includes the redoubtable Misery Ridge between Rockwood and Johnson Mountain.

The Seboomook forms a continuous valley on the southeast side of Misery Ridge, while the valleys of the Moose River and Misery Stream to the northwest of Misery Ridge have both formed within the Tomhegan. The northern end of Moosehead Lake also lies within the area of the Tomhegan and the Seboomook. The topographic expression of the rocks within this portion of the Moose River synclinorium is very reminiscent of the Valley and Ridge in Pennsylvania and Virginia, but on a smaller scale.

Figure 1. Generalized geologic map of the Greenville-Rockwood area. Numbers refer to field trip stops. Geology from mapping and compilation by Gary M. Boone, Syracuse University and Maine Geological Survey.



TABLE 1

DESCRIPTIONS OF ROCKS IN THE GREENVILLE-ROCKWOOD AREAS, MAINE

Map Symbol (Fig. 1)	Formation Name	Brief Description	Topographic Expression
		L A Y E R E D R O C K S	
Dto	Tomhegan	Fossiliferous dark metasandstone quartzite, slate	Low hills and valleys
Dtokm	Kineo member, Tomhegan formation	Rhyolite, tuff, flow breccia	Steep-sided hills, 200-800 ft. relief, (e.g. Mt. Kineo, Big Spencer Mtn).
Dt	Tarratine	Fossiliferous dark metasand-stone, metasiltstone, slate	Steep-sided ridges, (e.g. Misery Ridge).
Ds	Seboomook	Metasandstone and slate in thin graded couplets	Low hills except where subjected to contact Metamorphism, where major mountains are formed.
Dc	Carrabassett	Slate, minor metasandstone	Low hills except where contact metamorphosed to hornfels and granofels where high ridges and mountains are formed.
Sc	Capens	Red and green slate	Low hills
Su	Undifferen-tiated Sil-urian rocks		Low hills
Ov	Volcanics.in-cluding Kennebec and Lobster Mtn. volcanics	Felsite, tuff	Low to moderate hills
OCD	Dead River	Greenish metasiltstone and slate	Low hills
Ch	Hurricane Mountain	Rusty metasiltstone with exotic blocks	Low hills, valleys and basins. Exotic blocks form small knobs.
		P L U T O N I C R O C K S	
Dmp	Moxie	Tractolite, norite	Valleys, lake basins

Lobster Lake Anticlinorium

This belt forms a generally low area between Misery Ridge and the hornfels mountains and hills to the south. The headwaters of the Kennebec River has formed in the Dead River formation. The Hurricane Mountain formation underlies a broad area of low hills, while some small islands composed largely of exotic blocks occur in the southern terminous of the west outlet in Indian Pond (see Boone, this volume).

Only the Ordovician volcanics (Lobster Mountain volcanics) has extensive positive relief features, particularly toward the northeastern terminous of the structure.

Roach River Synclinorium

The Moxie pluton (see Hon, this volume) has played a major role in shaping the topography within this belt and to the east of Greenville in the mountains commonly called the White Cap Range (Hanson, this volume) in the headwaters of the East Branch and West Branch Pleasant River. The Moxie intruded the Carrabassett and Seboomook at a temperature estimated by Hon (this volume) to be in excess of 1100 C. The resulting thermal metamosphism removed the slately cleavage from the pelitic rocks and produced a resistant granofels. Most of the mountain ranges in Central Maine are granofels, produced by the contact metamorphism of previously low grade, regionally metamorphed, pelitic rocks. In southern Maine and Southern New England such contact metamorphism was less effective because higher grade regional metamorphism was experienced by rock further south, mineral assemblages were produced which were more stable when subsequently subjected to contact metamorphism and hence less altered. Mountains of granofels are practically non-existent in areas subjected to staurolite grade, or higher, regional metamorphism, except where intruded by very hot mafic rocks.

There are few mountains in Maine higher than the mountains surrounding with the Moxie pluton and similar plutons in the Sugarloaf Mountain-Bigelow Mountain area (Boone, 1973). Two prominent exceptions are Mount Katahdin and mountains in the Traveler Range. Katahdin owes its eminence to a resistant summit facies with a granophyric texture which protects the weak core of the pluton from weathering and erosion (Hon, 1980). Where the summit facies has been removed the highest elevations within the pluton are low hills. The Traveler rhyolite is comparable to the Kineo in compostion and resistance to erosion except that at present the Traveler is very much thicker and therefore forms higher mountains (Rankin, 1961).

The Moxie pluton underlies the lowest areas within the region. It forms a long, sinuous valley between the two hornfels mountain ranges. The southern end of the Moosehead Lake is underlain by the Moxie. The border of the Moxie, near its contact with the Carrabassett and Seboomook is fine grained and this rock weathers spheroidally. The coarser interior portion of the pluton undergoes granular disintergration and some areas of saprolite are preserved in this portion of the pluton. The contact phase of the Moxie is found near the summits of some high ridges, near Big Squaw Mountain, for example. The Moxie may have been protected from erosion by the hornfels rock on the northwest flank of the mountain or the finer grained Moxie may itself resist weathering and erosion.

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ROAD LOG

The Greenville, Moosehead Lake, and Brassua Lake quadrangles cover the area of this field trip.

Mileage begins at assembly point, Department of Transportation rest area 2.3 miles south of Greenville on Route 15.

Mileage

0 Rest Area, STOP 1 is about 0.1 miles south of rest area.

Stop 1 - Carrabassett formation. Slatey cleavage dominates the outcrop. Bedding is indistinct, but begins to show upon close inspection. Beds dip steeply northward and young in that direction. The Carrabassett here is typical of that found south of here in the Monson slate belt.

Return to cars.

2.5 Greenville, turn left at Indian Store.

3.7 Greenville Junction bear right on Route 15.

4.7 Stop 2 - Northeast end of granofels ridge of Carrabassett (?) Harfords Point township. Note prominent ridge on Figure 1 or on Greenville quadrangle called Little Squaw Mountain. This is the granofels ridge produced by the Moxie pluti that lies less than one-half mile to the west of here. Note the lack of slatey cleavage and that bedding is more pronounced than at the previous stop. A few small andalucite crystals are present. Walk to next outcrop.

4.9 Stop 3 - Little Squaw Mountain (T3 R5), Seboomook formation with prominent bedding, cross stratification and minor folds. On the southwest side of road toward Greenwood Motel is a greenish quartzite that may be the Madrid formation.

Return to cars.

5.3 Cross contact into the Moxie pluton.

5.4 Stop 4 - Fine grained troctolite (plagioclase, olivine). Well developed spherioridal weathering and development of iron-stained soil.

6.4 Stop 5 - Big Squaw Mountain Township (T2 R6). Coarse grained interior phase of Moxie. Note knots of mafic minerals which may result from crystal settling. Thi rock undergoes granular disintergration, with the development of grus. Above th jointed fresh rock is a zone of particularly decomposed rock that resemble pebbles, but consist of the knots of dark colored minerals surrounded by decomposed or disintergrated feldspar. Above this is a thick mantle of rusty soil composed of partially decomposed feldsfar crystals and clay. Although not typical of the clay-rich saprolites of the south, this material may in fact be a saprolite. We believe that the weathering represented here is probably pre-Wisconsin and may be partly pre-Pleislocene in development. The fact that any of this loose material escaped erosion by the ice sheets that passed this wa may mean that it was originally much thicker and also may have been frozen durin glaciation.

7.4 Last outcrop of Moxie on north west side of pluton. This rock resembles that seen at Stop 4. Total width of pluton here is about 2 miles.

7.8 STOP 6 - Spectacular outcrops of Seboomook. Thin (2-5 cm) graded beds with large andalucite crystals in the upper pelitic layer. This gives the impression that the pelitic layers are coarser and that the beds fine downward into the sandy layer. The beds in fact fine upward. These effects are best seen in the weathered upper surface of the outcrop where there are also small folds interpreted as slump phenomena. The rocks here are granofels, with a sparkling luster from the coarse grained biotite, andalusite and possibly cordierite. The outcrop is the northern end of the large ridge that ends on the sumit of Big Squaw Mountain. Before leaving, look back along the road to Greenville. Visible in the background is the hornfels ridge of Little Squaw Mountain and in between is the valley formed in weathered Moxie pluton.

8.0 Turn left into entrance of Squaw Mountain lodge for lunch and for Stop 7 at the summit of Big Squaw Mountain after lunch. The mileage for the remaining afternoon portion of this field trip will pick up at the entrance road as we will continue north on Route 15 into the Lobster Mountain and Moose River structural belts.

STOP 7 - Summit of Big Squaw Mountain beds of Seboomook hornfels rock. Fine-grained border rocks of Moxie lie a few hundred feet to the southwest. From this vantage point, the broad lowland of the Lobster Mountain anticlinorium is visible with Misery Ridge beyond and the Kineo trend beyond that. Slightly to the northeast are two long mountains. The easterly mountain is Big Spencer composed of Kineo volcanics. Far to the northeast is Katahdin.

Return to bottom of mountain and assemble in caravan at end of entrance road to Squaw Mountain.

8.0 Turn left on Route 15 toward Rockwood.

10.8 Undifferentiated Silurian rocks.

12.8 Capens formation

13.8 West outlet, headwaters of Kennebec. Optional Stop 8. Green phylitic slate of Dead River formation.

14.6 Hurricane Mountain formation.

20.6 STOP 9 - Tarratine formation at northeast end of Misery Ridge. Fossiliferous metasandstone, metasiltstone and metapelite with well developed slatey cleavage.

21.4 Look to right across narrows to Kineo Mountain. Kineo rocks are exposed in bushes on left.

22.6 Moose River and exposures of Tarratine are visible on far shore.

26.6 STOP 10 - Southwest termination of Kineo member occurs a few 10's feet beyond right side of road. Rhyolite displays prominent layering with slump features displayed.

END OF TRIP.

Proceed around sharp bend to wide gravel shoulder. Brassua Lake is visible. Turn around and return to Squaw Mountain for Annual Supper.

- THANK YOU -

TRIP C-3

Stratigraphy and Sedimentation in Silurian Flysch East of Millinocket, Maine

By

David C. Roy, Nancy A. Demorest, and Maureen J. Hill
Boston College

East of Millinocket is a broad belt of slate and graywacke that forms the northeastern extension of the Maine Slate Belt. Current stratigraphic studies in this belt of rocks by us in the northwestern part of the belt and by A. Ludman, of Queens College, farther southeast suggest fairly good lithologic correlations with the Sangerville-Smalls Falls-Madrid succession of west central Maine as described by Ludman (1977; 1979) and Pankiwskyj and others (1976). The purpose of the present trip is to illustrate the sedimentology of the Allsbury and Lawler Ridge formations (Sangerville and Madrid equivalents respectively) and to examine the gradational transition from Allsbury to Lawler Ridge through an un-named sulfidic phyllite unit (Smalls Falls equivalent?). In addition we will examine a large exposure of sheared/mylonitized Allsbury Formation(?) inferred to be along a fault.

Background

Regional Setting

As shown in Figure 1, the Allsbury and Lawler Ridge formations form a broad flysch belt which lies to the east and southeast of the Weeksboro-Lunksoos Lake Anticlinorium (Neuman, 1967; Ekren and Frischknecht, 1967) in which pre-Silurian rocks are extensively exposed. The eastern flank of the anticlinorium is broken by numerous faults and is in fault contact with the Allsbury Formation of the slate belt. The faults and anticlinorial core rocks are cut by the Katahdin Granite which was intruded late in the Acadian Orogeny.

The oldest rocks of the anticlinorium comprise the Cambrian (?) Grand Pitch Formation (Cgp) which is a diverse unit in which variegated slate and quartzite predominate. Large terranes of Middle Ordovician volcanic rocks (Ov) and chert (Ow) are present as apparent fault slivers in the portion of the anticlinorium shown in Figure 1. Along strike to the north in the anticlinorium, tuff and tuffaceous sedimentary rocks of the Early Ordovician Shin Pond Formation are also present. These Cambro-Ordovician units are cut by the Rockabema Quartz Diorite which was intruded in the latest Ordovician or earliest Silurian.

Also occurring in fault slices of the southeastern flank

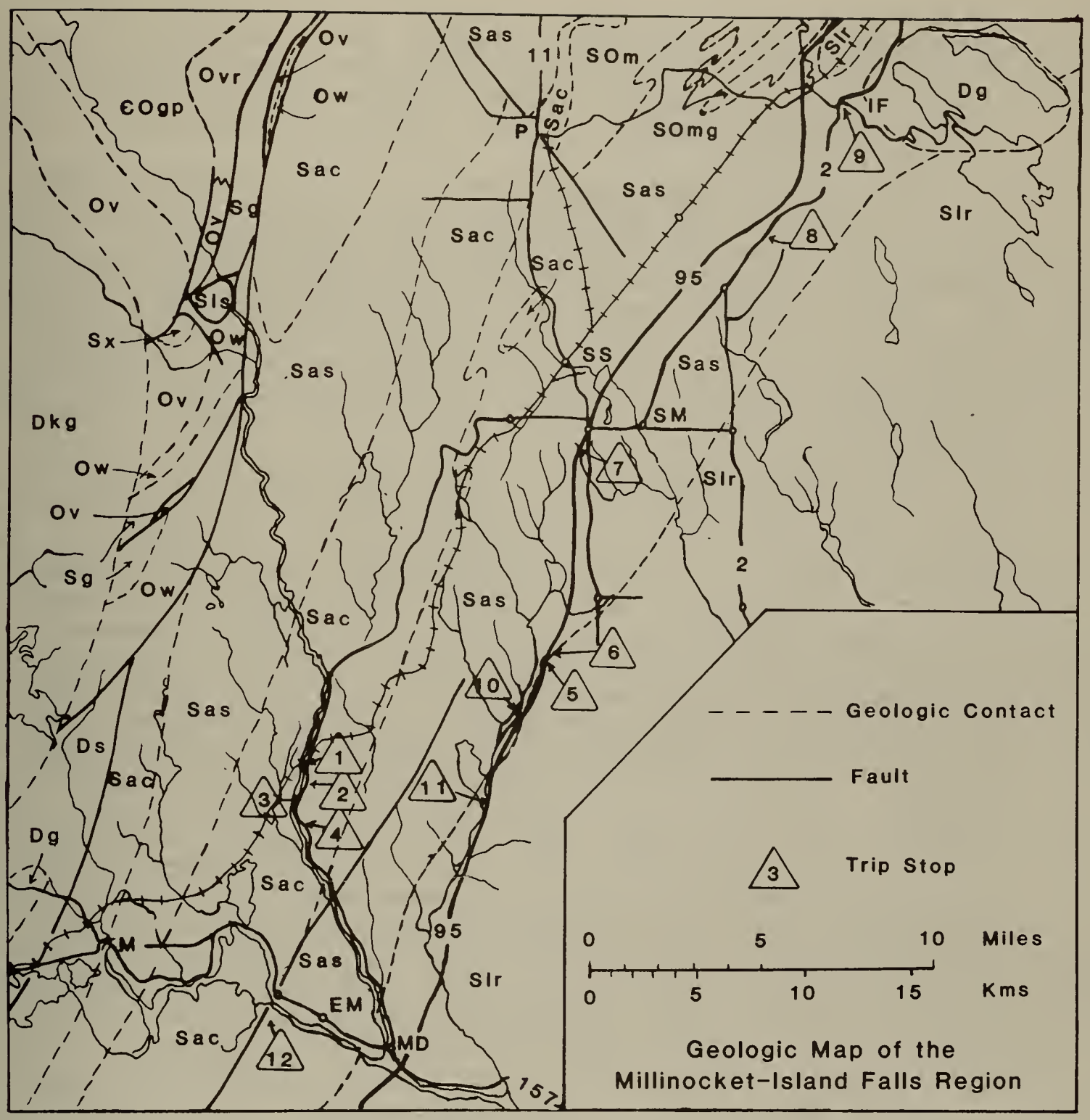


Figure 1: Generalized geologic map of the Millinocket-Island Falls region of northeastern Maine. Map is compiled fom maps produced by Neuman (1967), Ekren and Frischknecht (1967), and Roy (1981; unpublished data). Formational Units are: €Ogp, Grand Pitch Formation; OV, un-named volcanic rocks; Ovr, un-named volcanic rocks intruded by Rockabema Quartz Diorite; Ow, Wassataquoik Chert; SOm, Mattawamkeag Formation; SOmg, Carys Mills Formation of the Meduxnekeag Group; Sls, un-named limestone; Sx, un-named tuff breccia; Sg, un-named conglomerate, sandstone and slate; Sac, Sandstone Member, Allsbury Formation; Sas, Slate Member, Allsbury Formation; Slr, Lawler Ridge Formation; Ds, undifferentiated Devonian Slate; Dkg, Katahdin Granite; Dg, Devonian Granite. Cities and Towns: M, Millinocket; EM, East Millinocket; MD, Medway; SM, Sherman Mills; SS, Sherman Station; P, Patten; IF, Island Falls.

of the anticlinorium are Silurian rocks which consist of both Late Silurian limestone (Sls), Tuff breccia (Sx), and a succession of interbedded sandstone, conglomerate, and slate (Sg) that is in many ways lithologically similar to more sandstone-rich phases of the Allsbury Formation to the east (Neuman, 1967).

Along the southeastern margin of the Maine Slate Belt at the latitude of the trip area, but outside the map of Figure 1, are the rocks of the Miramachi Anticlinorium. A. Ludman (personal communication) has concluded the slate-belt rocks are in fault contact with the anticlinorial core rocks which consist of several Cambro-Ordovician units. The Cambro-Ordovician units are similar to those assigned to the Tetagouche Group in adjacent New Brunswick and they also have lithologic affinities to rocks of similar age in the Weeksboro-Lunksoos Lake Anticlinorium.

The Maine Slate Belt rocks seen on this trip therefore lie, and were probably deposited, between two terranes containing similar pre-Silurian rocks. In addition, the core rocks of both anticlinoria underwent Taconian deformation and became positive areas during the Silurian (Roy, 1980). Provenance determinations based on lithic fragments in the sandstone and conglomerate beds in the Silurian flysch, paleocurrents, and lithofacies variations along the western margin of the slate belt show clear derivation of detritus from a northwestern land area of which the core rocks of the Weeksboro-Lunksoos Lake Anticlinorium were a part (Roy and Mencher, 1976; Neuman, 1967; Roy 1980). This introduction of detritus from the west into the flysch basin probably began in the earliest Silurian (Ayrton and others, 1969), was certainly well established by the Lake Llandoverian of the Silurian, and continued into the Wenlockian with a substantial reduction in coarse clastic sediment. This reduction in coarse clastic sedimentation during the Early, Middle, and locally the Late Silurian has been attributed both to the reduction in source-area relief and to the transgression of the Late Silurian sea westward (Roy and Mencher, 1976; Roy, 1980).

The provenance relationships of the Silurian flysch sediments along the eastern margin of the basin to the uplifted Miramichi terrane is less well documented but seems to be broadly similar to that observed along the western margin. The similarity in core rocks of the two anticlinoria make it difficult to separate the source contributions of each to the flysch sandstones using only lithic fragment populations. Ultimately, detailed lithofacies analysis will have to be completed to adequately sort out the sediment dispersal patterns.

It appears likely that the Silurian flysch we will examine rests on a Late Ordovician and/or earliest Silurian sequence of micritic limestone, slate, and graywacke generally characteristic of the Aroostook-Matapedia belt. This sequence, the Meduxnekeag Group of Pavlides (1965) in Maine, underlies Silurian flysch (Smyrna Mills Formation) and forms an extensive terrane between Houlton and the southern Gaspe Peninsula.

South of the latitude of Island Falls the rocks of the Aroostook-Matapedia sequence are only questionably exposed at the surface and we lose touch with the bottom of the Silurian flysch sequence in the interior of the slate belt.

Stratigraphy within the Silurian flysch

The Silurian flysch east and northeast of Millinocket has been divided into two formations. The Allsbury Formation has been extended into the Millinocket region from the Island Falls Quadrangle (Ekren and Frischknecht, 1967) and the Stacyville Quadrangle (Neuman, 1967). The Allsbury is overlain conformably by the Lawler Ridge Formation, a newly defined unit (Roy, 1981). Along the contact between the formations is a thin un-named interval of sulfidic phyllite and sandstone which can so far be mapped only at large scale along Interstate 95 at and south of Lawler Ridge (Stops 5 and 10).

Regional metamorphism is at low greenschist grade (sub biotite) and sedimentary structures are generally well preserved. Graywacke beds are typically cleaved and the cleavage obscures the details of depositional structures in some beds. Our detail studies of the physical stratigraphy and petrology utilize excellent exposures along the East Branch of the Penobscot River in the Millinocket Quadrangle and along Interstate 95 near Salmon Stream Lake in the Mattawamkeag and Sherman Quadrangles.

Allsbury Formation

The Allsbury Formation was originally defined by Ekren and Frischknecht (1967) in the Island Falls Quadrangle and extended into the Shin Pond and Staceyville quadrangles by Neuman (1967). Roy (1981) carried the formation into the Sherman and Mattawamkeag quadrangles and concluded that the "Rocks of Island Falls" of Ekren and Frischknecht (1967) is lithologically synonymous. As pointed out by Neuman (1967) the formation is a monotonous succession of gray and green-gray phyllite with variably abundant thin "clean" siltstone beds and somewhat thicker quartzose sandstones. Dark gray-to-black phyllite seems to form horizons within the formation but these cannot be followed very far in outcrop but have been traced electromagnetically by Ekren and Frischknect in the Island Falls Quadrangle. Red and green slate also outcrop sporadically but also cannot be traced. Thick-to-massive bedded, cleaved and non-cleaved, quartzo-feldspathic and lithic sandstone are present in the Allsbury terrane and tend to be abundant in some places and absent elsewhere. Without marker horizons it is very difficult to subdivide the formation; a problem L. Pavlides faced while mapping the lithologically similar and largly coeval Smyrna Mills Formation along strike to the northeast. Neuman (1967) divided the formation into a sandstone-rich member (between 25 and 50 percent sandstone) and

a thin-bedded slate-dominated member. Neuman inferred that the sandstone-rich member is the older and possibly is in part correlative with pebble-rich sandstones that contain poorly preserved brachiopods of Llandoverian aspect; he assigned the fossiliferous sandstones to the Frenchville Formation which was mapped as immediately underlying the slate member of the Allsbury.

Roy (1981) attempted to follow Neuman's subdivision of the Allsbury terrane by mapping a Sandstone Member containing approximately 30 percent sandstone which commonly includes beds in excess of 30 cm. (1 foot) in thickness and locally has sandstone intervals with beds in excess of one meter. The type exposures for the Sandstone Member are along the East Branch of the Penobscot River between Grindstone Falls and Meadow Brook Rips (Stops 1, 2, 3 and 4). Roy's Slate Member contains over 75 percent gray and lesser green and red phyllite with variably abundant fine-grained sandstone and siltstone beds that rarely approach 30 centimeters in thickness. The Slate Member is well exposed along Interstate 95 between Sherman and Island Falls (Stops 6, 7, 8, and 9). In small pavement or roadcut exposures, sandstone forms over 70 percent of the outcrops where the Sandstone Member is present; slate/phyllite is predominant where the Slate Member is present. Estimates of the real proportions of rock types must be done in large stream exposures or road cuts.

The Sandstone Member is believed to generally underlie the Slate Member but it is likely that the contact as mapped in Figure 1 is not strickly an isochronous sedimentary boundary. In flysch sequences of this type, transitions from sandstone-rich to sandstone-poor facies take place both vertically and laterally; therefore, the Sandstone Member may be both a lower and western facies though other relationships are conceivable. If, for example, the formation developed as a submarine fan complex as we believe, then sandstone-rich sequences can be perched at various stratigraphic levels and at many separate positions along the trend of the paleoslope. In addition, the thicker and coarser sandstone bodies should be "shoe-string sands" if they formed in submarine fan channels as we suppose; such sand bodies would provide little help in tracing out folds because they are not of wide enough lateral extent.

Un-named Sulfidic Phyllite

A gradational contact is observed in exposures along Interstate 95 between the Slate Member of the Allsbury Formation and the overlying Lawler Ridge Formation (Stops 5 and 10). An un-named unit consisting of sulfidic and carbonaceous phyllite with thin rust-weathering parallel-laminated and ripple cross-laminated siltstone and fine-grained sandstone forms the transitional zone. Previously included as part of the Allsbury Formation, as one of its black slate horizons, this sulfidic unit is now being traced out along the

Allsbury-Lawler Ridge contact. Roy and Forbes (1970) reported on graptolites obtained from this unit at Stop 5 that W. B. N. Berry considered to be of Late Llandovery age. A. Ludman (personal communication, 1982) recovered graptolites of Early-to-Middle Wenlockian age from sulfidic slate he thinks belongs to the same transitional unit in the Smyrna Mills Quadrangle. An age near the Llandovery-Wenlock time boundary is probable for the un-named sulfidic phyllite.

Lawler Ridge Formation

Roy (1981) suggested that a sequence of massive quartzo-feldspathic graywacke with minor slate/phyllite in the Sherman and Mattawamkeag quadrangles be assigned to a new formation, the Lawler Ridge Formation. Ludman and Roy (unpublished data) have shown that the formation forms a northeast-trending 30-kilometer wide belt that extends northeastward into the Smyrna Mills and Houlton quadrangles and southwestward well beyond Medway. Ludman's work suggests that the Lawler Ridge belt may be continuous with the Madrid Formation with which it is lithologically similar.

The transition from the Slate Member of the Allsbury to the thick bedded sandstone sequence of the Lawler Ridge Formation suggests a rejuvenation of an old source area or the development of new sources. Paleocurrent data are as yet too fragmentary to establish changes in sediment transport directions. What seems to be clear is that, rather suddenly, large influxes of sand of fairly uniform grainsize took place producing thick sandstone beds with poorly developed sedimentary structures. This influx of sand apppears to have been widespread in the Maine Slate Belt and seems to follow a period of diminished sand deposition. The initiation of thick-bedded sand deposition represented by the Lawler Ridge is apparently of Wenlockian age; elswhere in the Maine Slate Belt massive sand deposition may have begun a little later. Roy (1981) has suggested that this influx of sand may have been initiated by uplift of eastern sources and not rejuvenation of the western "Taconia" source which in northern Maine, at least, appears to have been invaded by Late Silurian shallow-water seas.

Structure and Metamorphism

A nearly vertical foliation (S_1) penetrates pelitic and siltstone beds of all units of the Silurian flysch in the region. The cleavage in phyllite intervals is axial planar to upright, similar, highly appressed mesoscale F_1 folds seen in large outcrops (Stop 8) and larger F_1 folds as mapped along the East Branch of the Penobscot River (Stops 1, 2, and 3). Most sandstone beds are well foliated with cleavage commonly refracted with respect to the pelitic beds. Some massive sandstone beds in channel sequences of the Allsbury Formation

are not cleaved or are only very poorly cleaved. Regionally, F_1 folds plunge at angles of 20 to 90 degrees to the north east and southwest. Rare F_2 folds are seen in the region; they are generally steeply plunging low amplitude warps or open concentric folds. At present the regional importance of these later folds is not understood.

Metamorphism has carried the rocks to low greenschist grade; contact metamorphism near plutons caused hornfels of biotite grade or higher to develop. The regional metamorphic assemblage is quartz-albite-muscovite-chlorite ± carbonate ±pyrite. Porphroblasts of pyrite and ankerite are common and well displayed in many of the outcrops on Interstate 95.

Faults are seen in almost every large exposure in the region. Most of the faults are "slips" of uncertain displacement parallel to the foliation and appear to be sharp interfaces. There are zones of disruption of stratigraphic continuity in some exposures which may be related to more substantial fault displacements than those represented by the discrete faults just described. The disruption of stratigraphic and fold continuity occurs in zones from a few meters, as in the large exposure at the Sherman interchange on Interstate 95, to possibly a kilometer, as at and downstream of Dolby Dam (Stop 12). The disruption may involve only the fragmentation of the competent beds producing bed segments and delimbed fold hinges or, in the extreme, has produced mylonite. Most of this sort of disruption has been observed in rocks assigned to the Slate Member of the Allsbury Formation but it may well be that the entire slate belt is laced with anastomosing steeply dipping minor and major faults that are generally parallel to the structural grain of the region. The dimensions of the intervening zones of more or less coherent stratigraphy and fold patterns are important to understand, but as yet they not known.

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Itinerary

The trip will assemble at the rest area at Grindstone Falls located on Maine Route 11, nine miles north of its intersection with Maine Route 157 in Medway at 8:30 AM. Route 11 runs along the eastern bank of the East Branch of the Penobscot River from Medway to Grindstone.

Mileage

0.00 STOP 1: Allsbury Formation, Head of Grindstone

Falls, East Branch of the Penobscot River. Park in the rest area.

The Grindstone Falls section, which includes Stops 1-3 on Figure 2, displays an unusually thick sequence in continuous exposure of the Sandstone Member of the Allsbury Formation. Stop 4, at Meadow Brook Rips, is an isolated exposure that is separated from the outcrops at Stop 3 by about 2 kilometers of river that has very few exposures. Within the Grindstone section, structural deformation appears to be relatively simple, consisting of an anticline-syncline pair, the axes of which are shown in Figure 2. These folds are cylindrical with local plunges along their axes of no more than 25 degrees. Stratigraphy within this section is reminiscent of submarine fan models proposed by Walker (1978) and Mutti-Ricci Lucci (1972: as translated from the Italian by Nilsen 1978). They recognize the importance of thickening- and coarsening-upward sandstone sequences along with classical turbidites as indicators of outer fan deposition. Thick, coarse-grained sandstone and conglomerate sequences with little shale are seen as having been deposited within submarine channels. At Stop 1 we see fine- to medium-grained quartz- and feldspar-rich sandstone beds up to 55 cm. thick which are interlayered with thinly bedded (1-5 cm.) siltstone, micrite and shale (now phyllite). Although the micrite beds are a minor component, the rocks in the section are commonly calcareous with several sandstone beds containing ellipsoidal carbonate concretions up to 5 cm. in length. Approximately 30 percent of the Grindstone Falls section is composed of sandstone. A partial measured section of the sequence here is presented in Figure 3A which highlights the occurrence of the sandstone beds. A detailed description of the intervening pelite-rich intervals is shown in figure 3B. The sandstone beds seen here represent the finer-grained and thinner sandstone beds characteristic of deposition in a relatively distal, outer fan environment (Walker, 1978). Several cycles of sedimentation within this portion of the Grindstone Falls section can be defined using the occurrence of multiple, closely-spaced, or unusually thick sandstone beds to represent the tops of coarsening- and thickening-upward sequences. These cycles vary in thickness; encompassing between 1 and 10 meters of section from the pelite-rich base of the cycle to its highest sandstone bed and internally they contain numerous classical turbidites. Several of these thicker sandstone beds form the ledges which can be seen to cross the river

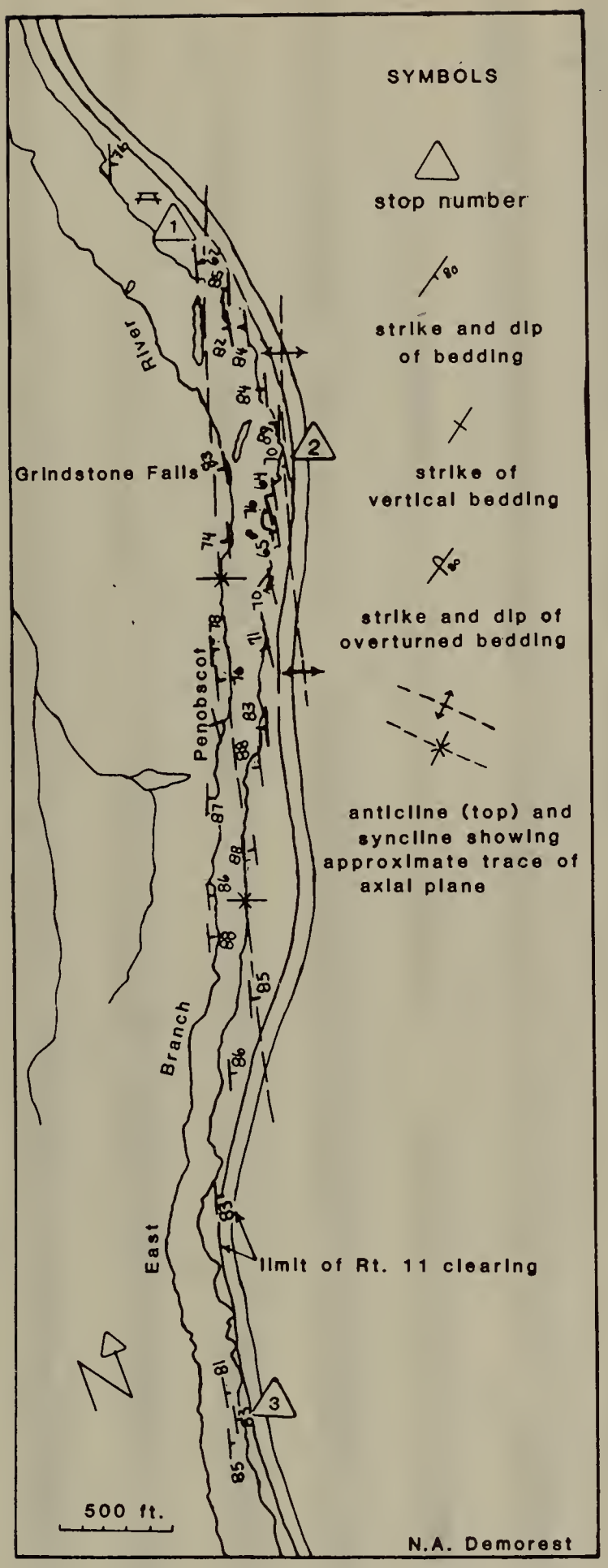


Figure 2: Detail geologic map of a portion of the East Branch of the Penobscot River at Grindstone Falls showing Stops 1, 2, and 3.

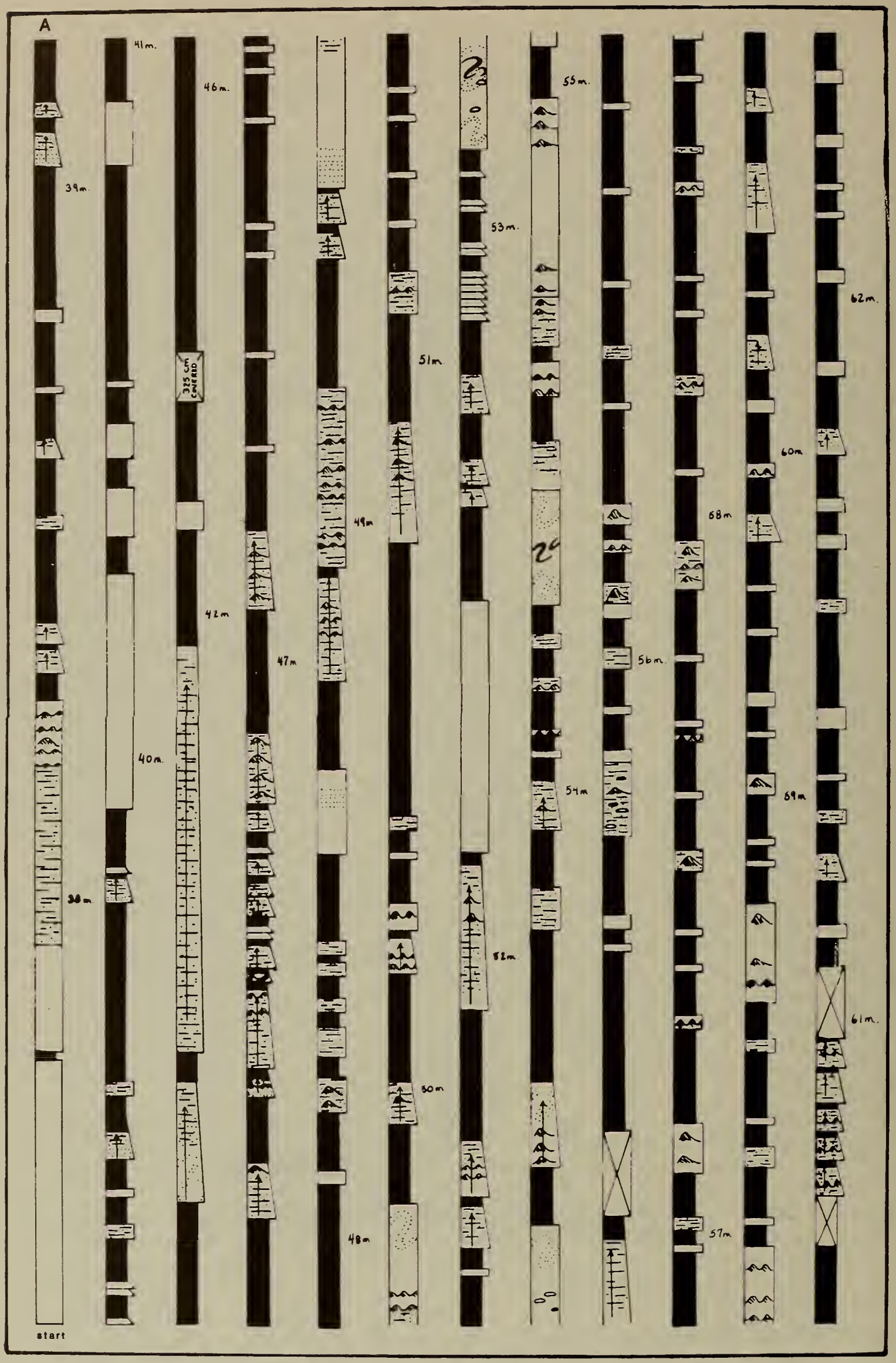


Figure 3: Partial stratigraphic section of the Sandstone Member of the Allsbury Formation at the Grindstone Rest Area (Stop 1). The meter marks indicate stratigraphic distance above base of measured section which is about 37.5 meters below the "start" position on A. B shows a detail of one of the "siltstone/shale" intervals of A.

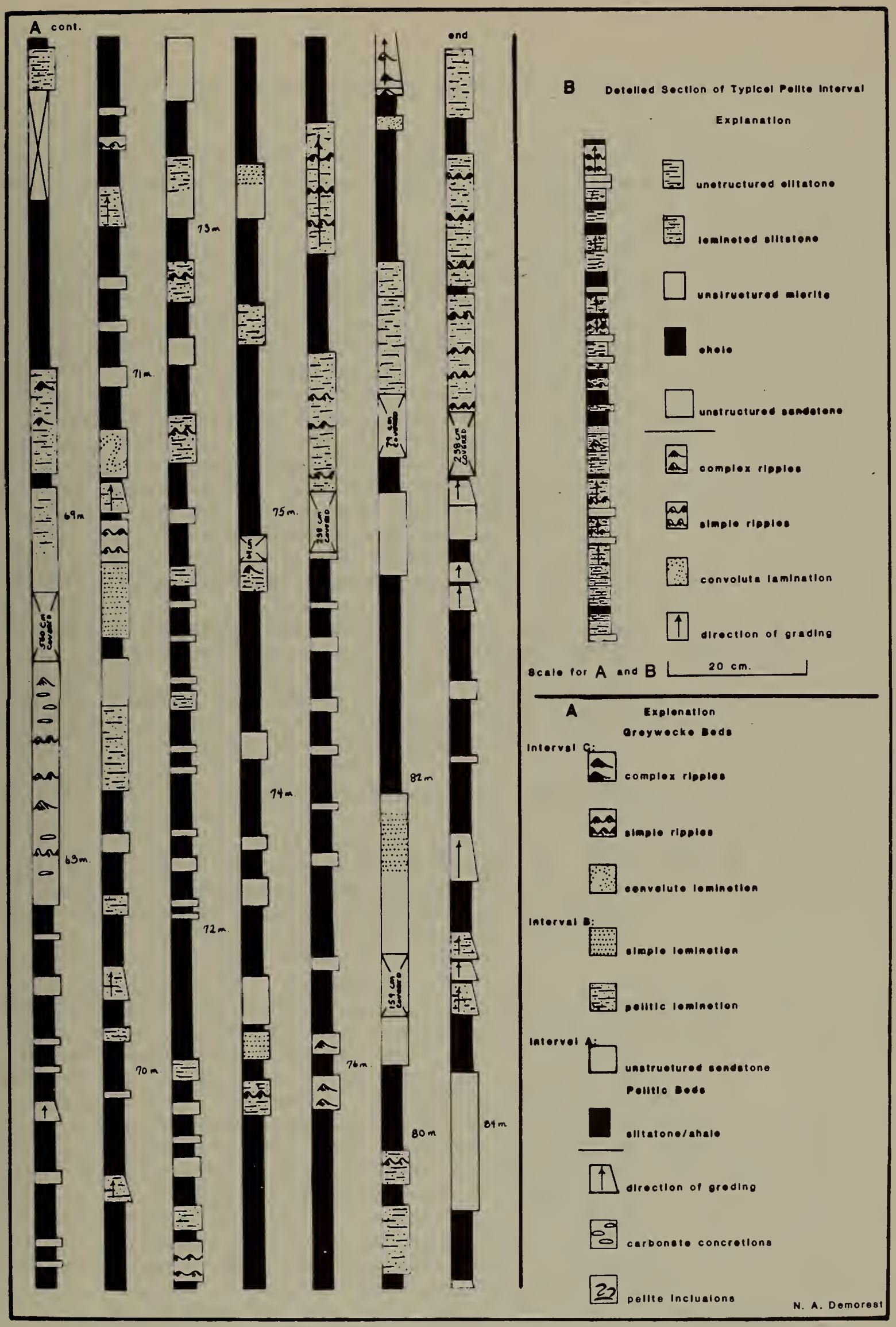


Figure 3 (Continued)

at the head of Grindstone Falls.

Ø.25 STOP 2: Allsbury Formation. This stop is located .2 miles south of the Grindstone Falls Rest Area, along Route 11, and may be approached easily on foot.

The section exposed here is stratigraphically below the section at Stop 1, and is involved in an anticline that plunges S16W at 23 degrees. This anticline is inferred to extend beyond the exposure as shown in Figure 2. The core of the anticline is composed of thinly cleaved pelite and thin sandstone beds similar to those seen at Stop 1. Incorporated on the west limb of the anticline however, is a 37 meter section of coarse-grained and largely unstructured sandstone beds which represent deposits in a submarine channel. The long axis of this channel may cross the strike of the Grindstone Falls section obliquely as suggested by paleocurrent measurements taken at this exposure. The precise lateral extent of these channel beds is obscured by the river and by overburden. It does not appear that these thick beds are present elswhere in the river section. It is inferred that the channel sandstones give way laterally on the west limb of the anticline to the thinner-bedded sandstone and pelite beds seen at Stop 1 and shown in Figure 3.

Continue south on Route 11.

Ø.85 STOP 3: Allsbury Formation at foot of Grindstone Falls. Park along the shoulder of Route 11. The exposure is located .1 mile downstream (south) of the widened portion of the shoulder.

Here is an exposure of relatively thick, coarse-grained sandstone beds typical of those deposited in submarine channels. Several beds thin laterally and are lensoid in shape which may be indicative of deposition in individual channels. It is also possible, however, that these lensoid sandstone bodies are of tectonic origin and are boudins. Some of the thick sandstone beds show lamination and one displays good flute marks. The position of these beds within the stratigraphic section is near that of the thinner and finer-grained sandstones beds depicted in Figure 3A and 3B, again suggesting the lateral inhomogeniety of this section.

2.3Ø STOP 4: Allsbury Formation, Meadow Brook Rips. When

parking along Route 11, be careful of the soft shoulder.

Here is the best example of thick submarine fan channel-sandstone beds exposed along the river. Several of these are amalgamated beds which total more than 2 meters in thickness and incorporate granule-size grains. In several instances it appears that the sandstone beds have intruded upward and pinched off intervening shale sequences, resulting in lozenge-shaped occurrences of shale between successive sandstone beds. The thin-bedded sequence that overlies the channel sandstones displays numerous isoclinal folds although the predominant direction of younging remains to the east. The stratigraphic relationship of this exposure to the Grindstone Falls section is uncertain because there is inadequate structural control to permit correlation into the section of rock we saw at Stops 1-3.

9.00 Intersection with Maine Route 157. Turn left (east).

9.80 Turn left (north) onto Interstate 95. Outcrops from here to Stop 5 are of the Lawler Ridge Formation.

13.35 Salmon Stream.

18.12 Exit right to scenic overlook.

18.67 Scenic Overlook: From here on a clear day it is possible to get a spectacular view of Mount Katahdin and its surrounding uplands. Below on both the north- and southbound lanes are outcrops of Lawler Ridge; Stop 11 is on the southbound lane. Continue north on the Interstate.

22.35 Herseytown Townline at mile 249.

22.90 STOP 5: Contact between the Allsbury and Lawler Ridge formations. Pull well off of the highway and park on the grass.

This long series of discontinuous exposures is the only place we have found where the un-named sulfidic phyllite unit can be seen in contact with both the overlying Lawler Ridge Formation and the underlying Allsbury Formation. At the south end of the exposure massive, orange weathering, light gray, quartzo-feldspathic graywacke beds of the Lawler Ridge Formation can be seen in contact with the thinly interbedded, black, finely cleaved pyritiferous phyllite and fine grained sandstone and siltstone. This contact is directly across from the

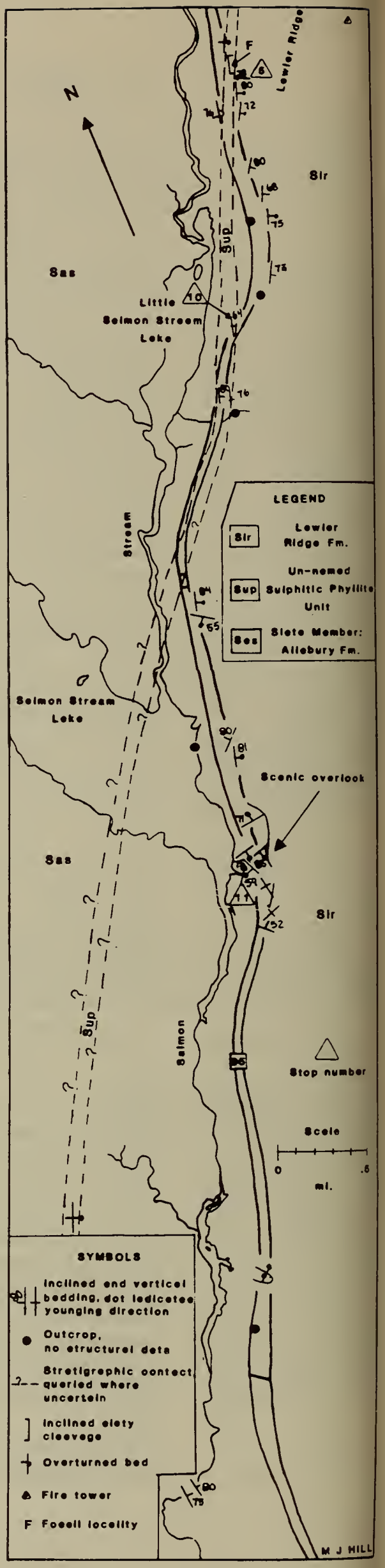


Figure 4: Detail geologic map of a portion of the contact between the Slate Member of the Allsbury Formation and the Lawler Ridge Formation along Interstate 95, showing Stops 5, 10 and 11. Note the Un-named Sulfidic Phyllite Unit located along the contact between the formations. Base is constructed from uncorrected air photos and orthophoto quadrangle maps of the Sherman, Millinocket and Mattawamkeag quadrangles.

speed limit sign and partially covered by bushes. The beds strike N15-20E and dip 75-85SE. The road cuts these beds obliquely at an angle of 7-12 degrees. The facing direction for this series of outcrops, everywhere we have been able to measure it, is to the southeast. The evidence for facing is provided by ripple cross-laminations as well as ripple bed forms in the sulfidic phyllite; facing in the Lawler Ridge Formation is indicated by sole features. Graptolites of probable Llandoverian age from this exposure have been described by Roy and Forbes (1970). Continuing downsection (north), approaching the guard rail along the right side of the Interstate, you begin to see bands of green-gray phyllite interlayered with the black phyllite, in pavements along the embankment. The mix of these two lithologies appears to be a transition to the Allsbury Formation. At the brook, small pavements and exposures can be seen that are typical of the Slate Member of the Allsbury Formation. Looking ahead up the road, a large exposure of this member (Stop 6) can be seen.

23.2 STOP 6: Allsbury Formation. Pull well off of the highway and park on the grass.

This is a typical exposure of the Slate Member of the Allsbury Formation. The rocks consist of gray phyllite thinly interlayered with orange-weathering fine-grained quartzose sandstone and siltstone beds. The phyllite makes up over 75 percent of this exposure. Ankerite and pyrite porphyroblasts are common here.

25.4 Casey Road and exposures of the Slate Member of the Allsbury Formation.

28.75 STOP 7: Allsbury Formation. Pull well off of the highway and park on the grass.

This is the best exposure of red and green slates that are typically found in the Slate Member of the Allsbury Formation. The green slate at the north end of the exposure displays nice grazing-trail trace fossils that have been getting harder to find in recent years as collectors have made off with them. A tightly appressed F_1 anticlinal hinge plunging to the southwest can be seen in red/purple slate at about mid-outcrop.

30.35 Sherman exit ramp.

35.45 STOP 8: Allsbury Formation. Pull completely off the pavement and park on the grass.

This exposure displays a mesoscale doubling plunging F_1 antiform which plunges N35E at 30 degrees and S35W at 16 degrees. The face of the roadcut is parallel to S_1 and passes through the fold very near the axial plane. The rock here is a gray silty pyritic slate that weathers to gray- and rust-colored layers. In fresh surface the rust-weathered layers are a lighter gray than the layers that remain gray during weathering; otherwise the layers are quite similar. Thin siltstone laminae are also common but none have proven convincing as facing indicators. It is of interest to determine clear facing in the slate sequence here to establish whether or not the fold is a straight-forward anticline.

36.95 Crystal Road.

41.00 Belvedere Road.

42.05 Exit to Island Falls.

42.35 Turn left (east) onto Maine Route 159.

42.75 Cross the West Branch of the Mattawamkeag River.

42.85 Bear right.

43.05 Junction of Route 2 in Island Falls. Continue straight (north).

43.10 Park across the street from J.D.'s Diner.

STOP 9: Allsbury Formation at Island Falls on the West Branch of the Mattawamkeag River in the town of Island Falls. This is our planned LUNCH STOP so carry lunches down behind J.D.'s Diner and across the old dam to the outcrops on the island in the river. Those who do not have lunch with them can get something at the diner or the store adjacent to the parking lot. To stay on schedule we can not stay here long so eat and look at the rocks at the same time.

This is the type exposure of the "Rocks of Island Falls" as described by Ekren and Frischknecht (1967). Roy (1981) included these rocks in the Slate Member of the Allsbury Formation as he mapped it in the Sherman Quadrangle. Thinly interbedded green-gray phyllite and laminated and cross-laminated siltstone and fine-grained sandstone are typical of the Slate Member in this belt of rocks we have been following since we crossed the

contact at Stop 5. These rocks are also quite similar to the Smyrna Mills Formation just to the northeast of here.

43.15 Continue straight at intersection with 159.

43.30 Bear left on Maine Route 159.

43.45 Cross the West Branch of the Mattawamkeag River.

43.95 Turn left onto Interstate 95 southbound.

45.35 Belvedere Road.

49.35 Crystal Road.

55.45 Sherman exit ramp, continue southbound.

60.95 Casey Road.

64.00 T1R6 town line.

64.65 STOP 10: Park on shoulder of highway, pull well off pavement; outcrop is on the left (east) side of the road.

This is the best exposure of the contact between the Lawler Ridge Formation and the un-named sulfidic phyllite unit. It is also an excellent exposure of the un-named sequence. The siltstones here display a rusty weathering due to disseminated pyrite and ankerite(?). Ripple trains and climbing ripples are common in the phyllite unit. Note the unstructured nature of the Lawler Ridge sandstone. Figure 5 shows the details of the stratigraphic section here. The sulphidic phyllite has been interpreted as being deposited in a reducing environment. The contact is, in this exposure, at the last appearance of the black phyllite and can be seen at the north end of the outcrop. There is some soft sediment deformation within the sulfidic phyllite at the contact. Shear zones are present throughout the section accompanied by quartz veins.

67.70 STOP 11: Lawler Ridge Formation. Park on the shoulder of the highway, well off the pavement; Outcrop is on the left (east) side of the road.

This is one of the better exposures of the Lawler Ridge Formation. It is located just west of the scenic overlook we stopped at earlier. Thick-bedded graywacke turbidites, up to about 4.5 meters are displayed here. These are best seen in the north end of the outcrop. Flame structures are common at

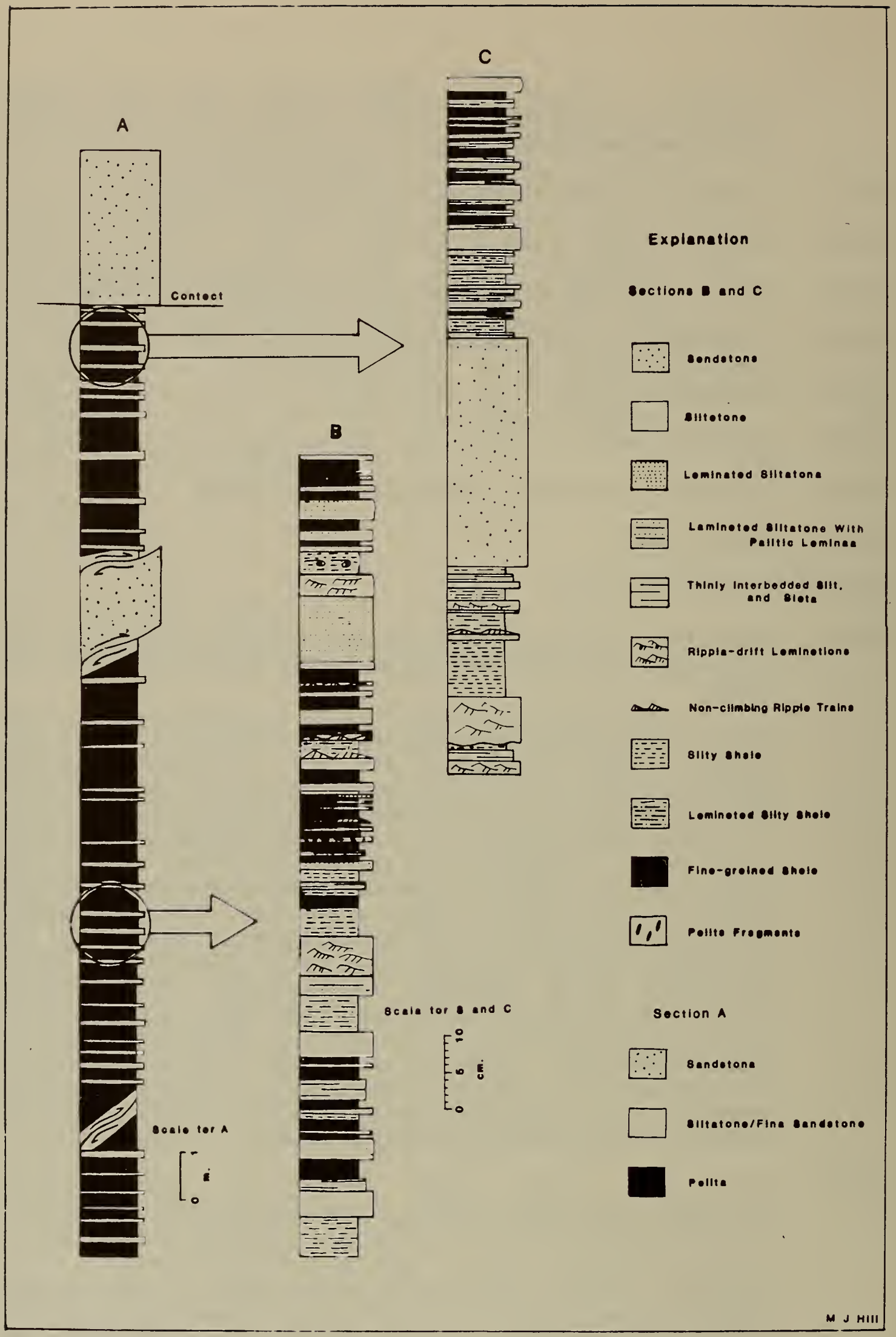


Figure 5: Stratigraphic columns for the section exposed at Stop 10. Generalized section A shows the contact between the Lawler Ridge Formation and the Un-named Sulfidic Phyllite Unit. The thickness of the siltstone and fine sandstone beds in the un-named phyllite unit are not to scale. Sections B and C are details of portions of the un-named phyllite unit.

the bottoms of the beds. Carbonate concretions and pelite clasts are also abundant. Note the pelite-clast conglomerate at the north end of the outcrop. Figure 6 shows the details of the stratigraphic section here. The strata top to the south and no reversals of facing have been found although there is a strike change about midway in the outcrop. Sole features may be seen on some beds, especially in the northern end of the outcrop.

73.05 Salmon Brook.

75.35 Exit to Millinocket-Medway.

75.50 Turn right onto Maine Route 157 west.

76.15 Junction of Route 11. Continue straight.

76.35 Cross the East Branch of the Penobscot River.

76.85 Take a left onto Route 116 south.

77.45 Cross the West Branch of the Penobscot River.

77.65 Take a right onto the un-named paved road.

77.80 East Branch, Medunkeunk Stream.

78.65 Park at the end of the pavement and consolidate cars. Take a right (north) onto the gravel road.

79.05 Turn left and drive along the power line.

81.70 STOP 12: Sheared and Mylonitized Allsbury Formation (?) at Dolby Dam on the West Branch of the Penobscot River. Park in the cleared area in such a way that we can easily head back out the way we came in.

This very large pavement is sketched in Figure 8. The pavement displays what is believed to be tectonically disrupted Allsbury Formation. It is possible to find portions of this exposure (points marked A) where the bedded character of the original stratigraphic sequence is fairly well preserved. Most of the surface shows pretty complete disruption of the stratigraphic and fold continuity leaving sandstone bed segments and delimbed fold hinges. Zones of mylonite (at B for example) passed through the disrupted formation and have abrupt, but gradational, contacts with it. The mylonite zones presumably represent zones of substantial shear displacements in which only small sandstone fragments survived. At C is a large synclinal fold hinge that has survived the disruption. As we

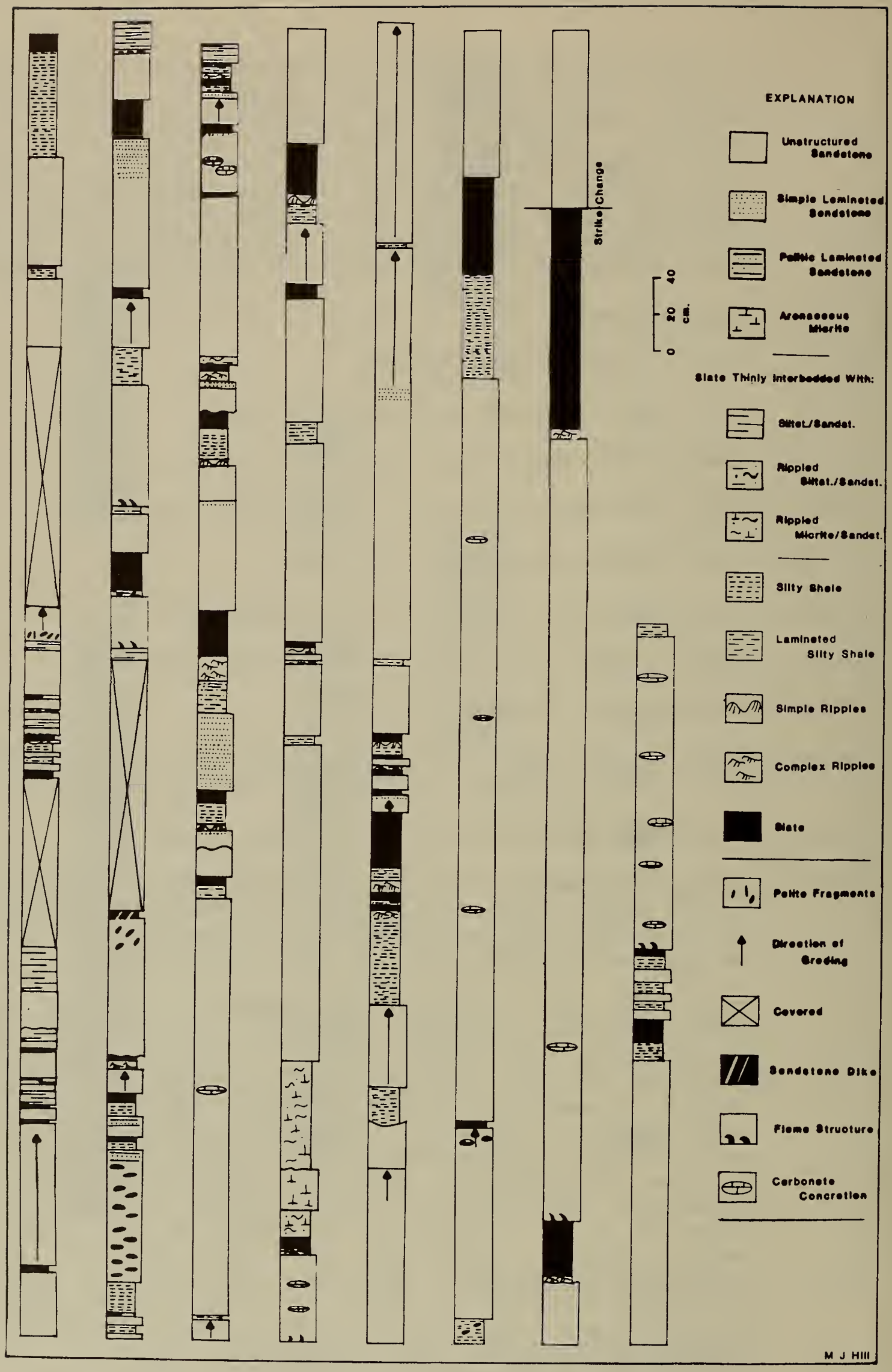


Figure 6: Stratigraphic column of the Lawler Ridge section at Stop 11.

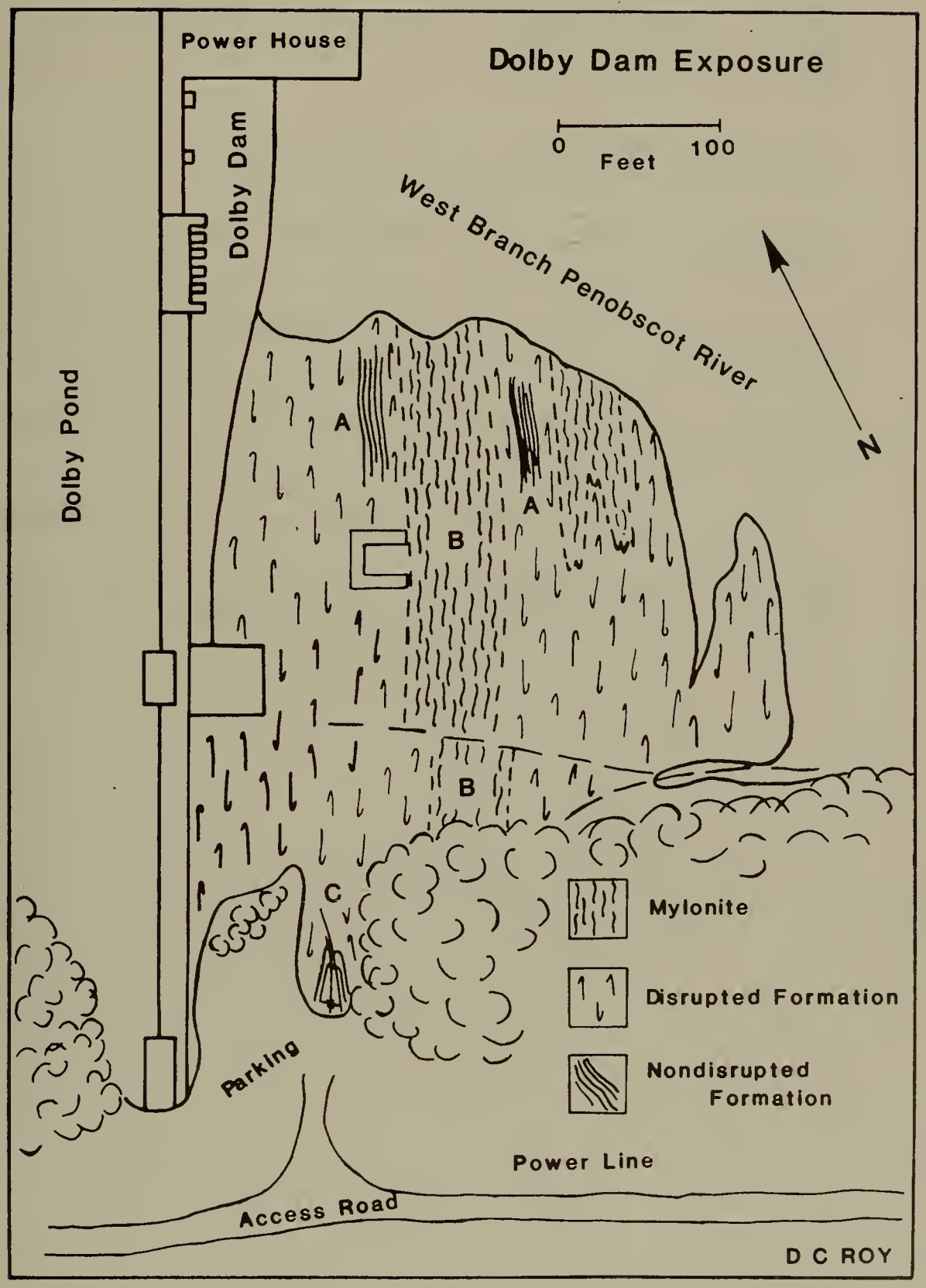


Figure 7: Sketch map of the large outcrop surface at the south end of Dolby Dam on the West Branch of the Penobscot River. Lettered areas are described in the text.

return along the power line we will pass over bedrock pavements showing disruption such as seen here, with intervening pavements showing more coherent patterns.

82.45 Bedrock pavement showing coherent folds in thin-bedded phyllite and siltstone beds.

82.70 Bedrock pavement of disrupted formation as seen at Stop 12.

84.80 Reassemble into cars. Continue east on the pavement.

85.80 Turn left at the triangular intersection onto Main Route 116.

85.95 Cross the West Branch of the Penobscot River.

86.55 Junction with Route 157/11. This is the end of the trip. Turn left to go to Millinocket and Greenville.

THE SEBOOMOOK LAKE AREA BECAME ICE FREE; BUT HOW ?

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INTRODUCTION

During the summer of 1983 we conducted surficial mapping and investigations in the Seboomook Lake area and along the West Branch of the Penobscot River. Reconnaissance checks and examination of prior work (Brewer and Caldwell, 1975) led us to hypothesize the existence of very large post- or proglacial lakes that spread from Moosehead Lake to Chesuncook Lake (Fig. 1). Low drainage divides now connect the two basins. The area is also interesting because at least 3 major river basins drain from it; the St. John River to the north, the West Branch of the Penobscot to the east and the Kennebec to the south. Additionally the headwaters of the Allagash Basin is only 50 km to the northeast. However one must drop 40 m from Seboomook Lake to that drainage. The extremely low topography provides numerous spillways for drift or ice-dammed lakes.

The region could possess complex deglaciation patterns because it lies just south of regions Kite and Lowell (1982) suggested as the site of the last remnant of a receding ice mass. Lowell (in press) documented the southward retreat of ice from the Maine-Quebec border. This contrasts with numerous works describing the northward recession of the same ice mass (e.g. Caldwell, 1975; Newman, 1982). Because the Seboomook region lies at the north end of previously mapped areas and just south of the ice divide, insight to the deglaciation of a small ice mass may lie waiting in the deep woods.

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The prior surficial mapping in the area is limited, neither Stone (1899) or Leavitt and Perkins (1935) got this far north. However, Caldwell (1975) first suggested that the Lobster basin (Fig. 1) contained a glacial lake dammed near Smith Halfway House on the West Branch of the Penobscot. He postulated two lake levels, the first at 1040 and the second at 1020+ feet, both draining into the Moosehead Lake basin. Glacial retreat to the north provided gravel deposits that covered the lake sediments.

The bedrock in the Seboomook area includes rocks of Lower Devonian to Ordovician or Silurian age. The map units (Marvinney, 1983) include the Seboomook Formation of cyclically bedded gray slate and fine sandstone; the Ironbound Mountain Slate of gray slate with minor greenish-gray phyllite; Frontenac Formation of interbedded gray to bluish-gray fine sandstone and wackie and thinly bedded dark slate; and the Canada Falls Volcanic Member of dark green basalt and minor intercalated dark green slate. Minor intrusive rocks include metadiabase and felsic dikes. The regional bedrock structure suggest these units continue northeast into the St John Pond and Northeast Carry quadrangles. However, Pollock (unpub.) mapping in quadrangles to the north, reports a somewhat different stratigraphy. In addition to the Frontenac and Seboomook Formations, Pollock includes green slate, basalt lava flows of low metamorphic grade, volcanic breccias and ashes, and sandstone and dark slate.

Given this setting we set out to map the distribution of lake sediments and deltas, measure the elevations of deltas, beaches and spillways, describe related sand and gravel deposits, and map other surficial features. Of course life is not that simple and what we found does not resemble the gravel deposits of southern New England. The drift sequence, although displaying all of the classic glacial features is quite thin.

On this trip we would like to focus on the deglaciation in the Seboomook Lake area. In this context three secondary themes fall out; 1) ice-flow history (one must consider of the interplay of Laurentide versus local ice); 2) deglaciation style of an ice mass of limited extent; and 3) postglacial lake development. Naturally we can not see the local evidence in any semblance of order, so we ask the reader to bear with us and shift his or her viewpoint back and forth from ice cap to gravel pit. Your input will aid preparation of the final report.

We wish to thank the following for assistance, both in the field and with data reduction; Rob Pockaly, Bill Naughton, Linda Healy, and Chris Viani. Discussions with W. Holland, W. Thompson and P. Calkin helped evolve our ideas. The generosity of Walter Anderson and the Maine Geological Survey provided the money and support to get into this mess in the first place.

ICE-FLOW HISTORY

In order to study the glacial events prior to deglaciation it is necessary to understand the ice-flow history. This information is extracted from striations etched onto bedrock surfaces. In the present study area approximately 250 observed locations show evidence of former ice-flow. Over 100 of these display multiple striations that demonstrate changes in glacier flow directions.

On an outcrop basis it is necessary to separate scratches which have associated azimuth criteria, such as stoss-and-lee forms or rat-tails, versus those which only show trends. This discrimination leaves 56 % of the outcrops with directional criteria. Because all azimuth locations show flow toward the E or SE quadrants, it is assumed that the trend localities also flow in the same direction. This assumption is made to allow a larger data base for relative age determinations. Care should be excised because regions to the north show a more complex history of flow reversals (Lowell, 1980).

When plotted on a Rose diagram (Fig. 2) 3 primary directions break out; E-ESE (50°-110°), SE (145°), SSE (>155°). These groups may seem somewhat arbitrary but are based on age relationships, map distribution, and relative erosional strengths.

Isolating only those multi-direction locations allows changes of ice-flow to be deduced. The relative ages are based on outcrop observations that younger striations can be traced up to, into, and across the floor of the older striations and grooves cut by prior ice-flow. Using the 3 different flow directions allows the following shifts to be noted:

1) from E-ESE to SE (31 locations)
2) from SE to SSE (8 locations)
3) from SE to E (29 locations)

This leads to the following ice-flow history. First ice flowed to the E-ESE with a major subsequent shift to the SE. The SE is the dominate flow which has a reduced shift to the SSE over-printed on it. The final flow is again to the E-ESE. The two flows to the E-ESE are separated by the stratigraphic position of the SE flow. The dominate SE indicators lie both above and below eastward flow indicators. A second possible explanation of the observed field data requires a single eastward flow with two SE flows, one below and one above. However, the first sequence draws support from the observation that those E indicators lying below the SE are oriented uniformly toward 115, are widespread, and are always grooves or large striations. In contrast the E indicators lying above the SE are variable with directions from 50 to 125°, are limited in distribution to west facing slopes, and are always faint etchings. At 3 locations the entire sequence of E to SE and back to E are present. The last eastward ice-flow has some strength because flow indicators are well displayed at high elevations on Little Russell Mt. in the St. John Pond quadrangle.

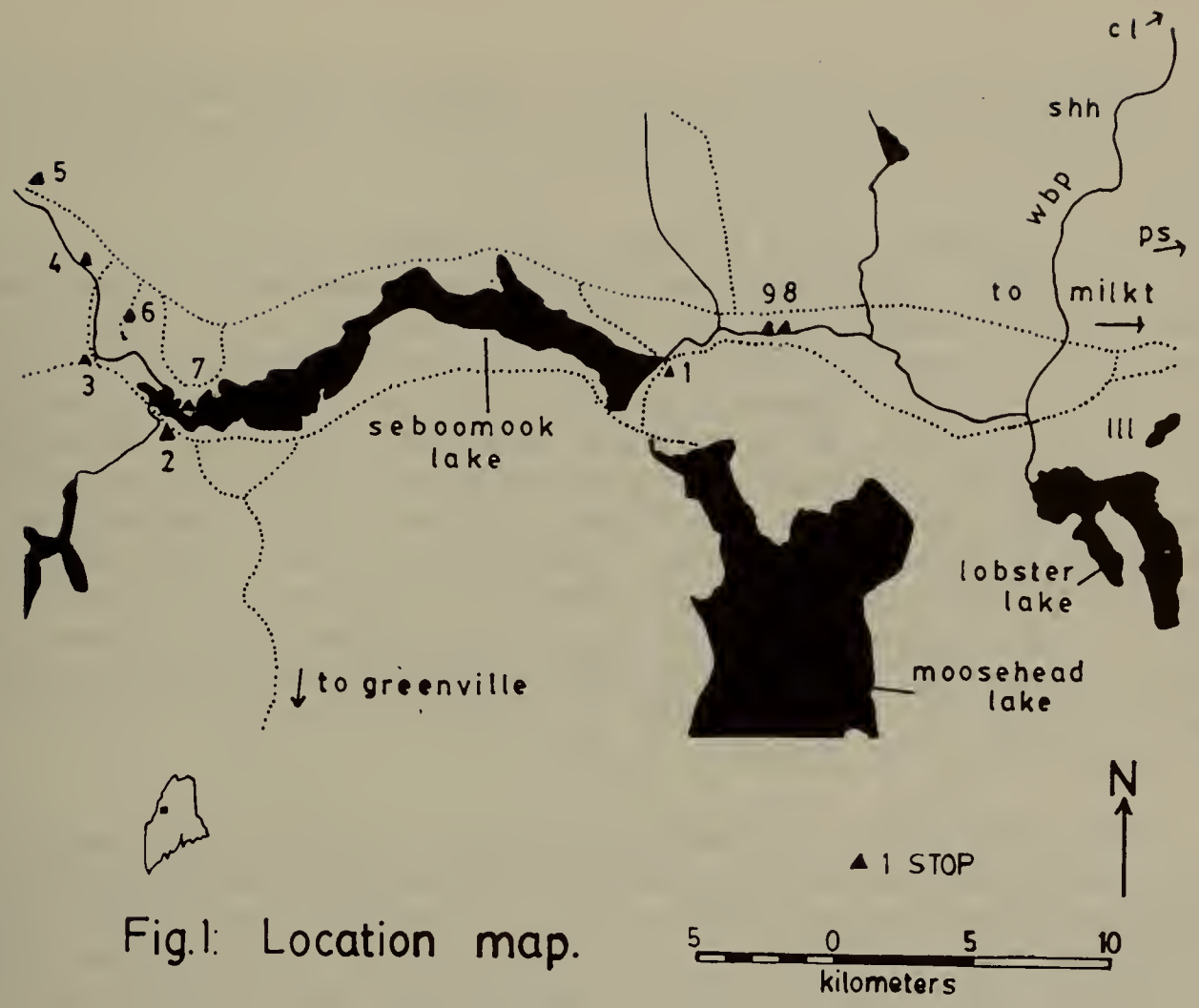


Fig.1: Location map.

Figure 1. Location map of the Seboomook Lake area. Stops shown as triangles. cl = Chesuncook Lake, shh = Smith Halfway House, ps = Pine Stream, lll = Little Lobster Lake, wbp = West Branch of the Penobscot River.

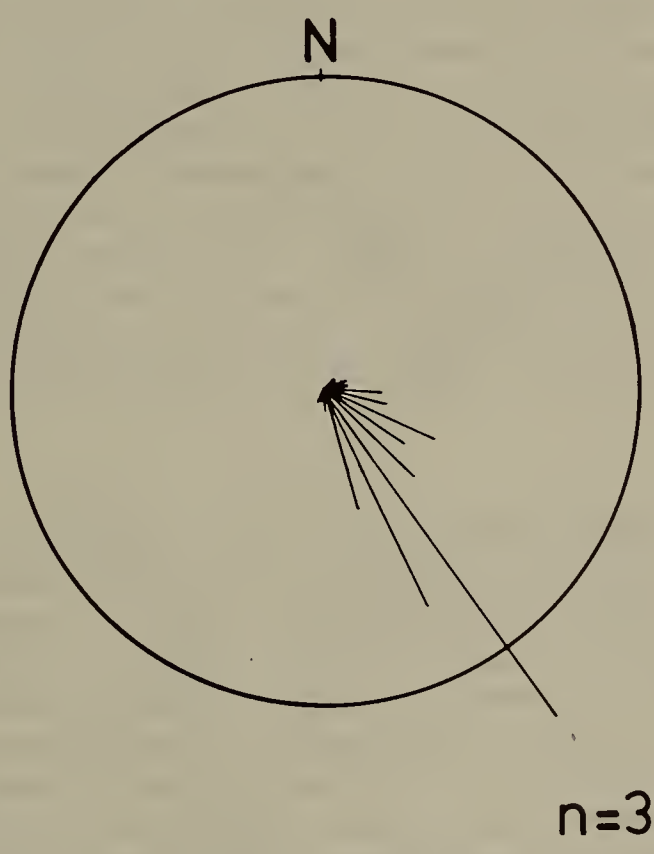


Figure 2. Rose diagram of striation locatilies. Circle at 25 %.

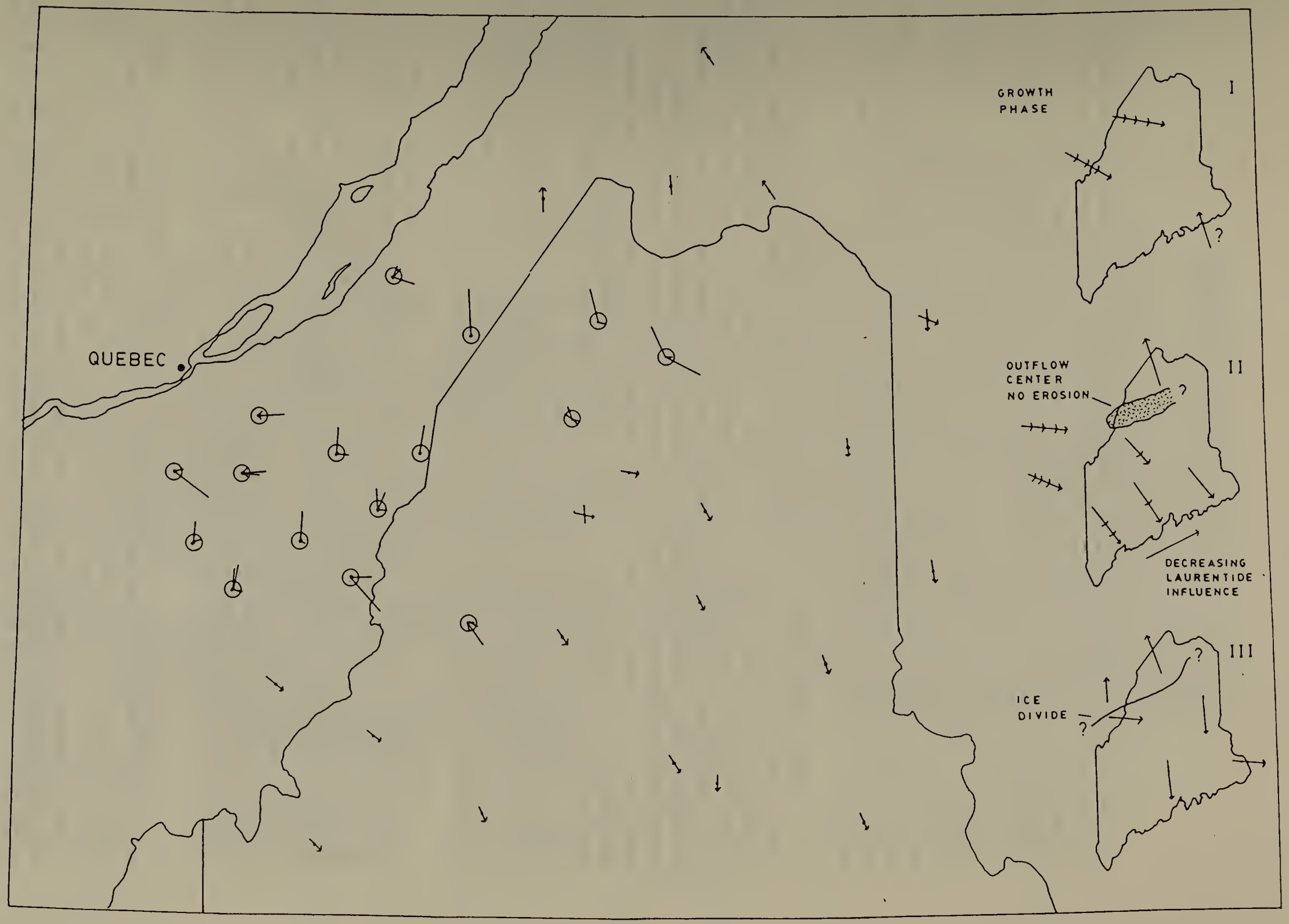
The extent of the last flow is presently unknown due to limited data from the east of this field area in the Chensuncook Lake region.

Placing this into a regional setting is a bit of a trick. LaSalle (1977) reported the following sequence from the region between the present study area and Quebec City. The first ice, Laurentide, advanced across the region toward 115-135°. Subsequently near Quebec City an easterly system developed during deglaciation. The relationship with the NE flow, also related to deglaciation is unknown. The very last flow, representing separated ice movement northward shows local deflections in its flow. Thus LaSalle reports a sequence of SE, E/NE and N ice-flows. In contrast, Shilts (1981) reported only the strong SE flow. Visual examination of his maps shows striations patterns similar to those of LaSalle for adjacent regions and this report, however Shilts did not break out the variations within the SE flow (Fig. 3).

The often used assumption, that Laurentide ice flooded all of New England the entire area with SE moving ice must be questioned because evidence for it can not be traced much further north than the Lac Megantic or Seboomook Lake areas. One suggestion extended for discussion, and it should cause some, is that the widespread SE flow results from an influx of Laurentide ice combined with outflow from northern Maine. As one moves from southwestern Maine along the coast toward the northeast the effects of Laurentide ice decrease and the effects of local ice increase (Fig. 3).

The ice stream moving through the St. Lawrence lowland may have persisted and drained both local and Laurentide ice through out the last glaciation. Laurentide ice, at times, could cross the St. Lawrence River, but its influence in northern Maine would be much less than in areas southwest toward New Hampshire and Vermont. Following this suggestion, the northward flowing ice (LaSalle, 1977; Lowell, 1980) may have occurred for most of the late Wisconsin. Although a continuous ice cover extended from Quebec to the Gulf of Maine, the ice interacted in response to more local conditions than as an overwhelming ice mass.

Figure 3. (overleaf) Regional ice-flow data. The large diagram shows the distribution of ice-flow data compiled from the following sources; LaSalle, 1977; Gauthier, 1980; Kite and others, 1982; Lowell, 1980; Halter, unpublished; Newman, 1980; Shilts, 1981; Hyland, 1981, Rampton and Paradis, 1980. The small circles represent rose diagrams with the circle at 10 %. The simple striation symbols represent visual estimates from that area. The right hand portion of the diagram shows postuladed interplay of Laurentide and local ice during the late Wisconsin. More hatchers on flow arrows indicate more influence of Laurentide ice.


QUEBEC
GROWTH PHASE
I
?
OUTFLOW CENTER NO EROSION
II
?
DECREASING LAURENTIDE INFLUENCE
ICE DIVIDE
?
III

MANNER OF DEGLACIATION

As the last ice moved eastward across the area it did little to influence the manner of deglaciation. Meltwater channels are the primary indicator of former ice position. Based on their distribution and character, our feeling is that the primary deglaciation style is downwasting. Largely inactive ice blocks separated and melted. Topography played the dominate role in controlling the distribution of the blocks and their drainage routes. The final deglaciation ties closely into the formation of proglacial lakes. Moreover this phase provides a wealth of stratigraphic information. Therefore the lake history follows at what might seem disproportional length. In reality the lake basins record only the final phase of deglaciation that began when the mass balance became negative.

The numerous channels spread over the study area fall into the classification of Derbyshire (1962). These include: marginal channels formed along the former ice margin, submarginal channels which record meltwater moving from a lateral position into the ice, subglacial channels which originated as the water moved through the base of the ice, and col gullies which occupy low passes in the hills and conducted meltwater from either subglacial conduits or ponded water spilling over a divide. The abundance of channels in this region compared to areas further north may result from a combination of the following factors; 1) high relief to record the downwasting, 2) thicker drift to record the meltwater erosion, and 3) more abundant meltwater supply. Over 65 channels were mapped in Seboomook Lake basin alone. Most of these are of submarginal or subglacial origin. Around the perimeter of the basin, col channels grace the hills.

Meltwater probably began cutting into the topography early during deglaciation (Fig. 4A). The highest channels west of Seboomook Mountain suggest temporary drainage to the north before erosion of a col channel to the south that now connects the Seboomook basin to the St. John drainage. Several esker segments north of the col lead into the channel which drained down to an elevation of 518 m (1700 ft) in the Seboomook basin. At this level and lower the drainage apparently moved southward across the basin to exit around several hills where channels conducted water at elevations from 430 m (1400 ft) to 362 m (1190 ft).

A major shift occurs as the ice level dropped and the meltwater began to move eastward across the southern flank of the basin (Fig. 4b). By this phase the ice seems to have lost much of its continuity because many channels at this level mark a shift from lateral to subglacial drainage positions. A particularly well developed system (Stop 6) cuts into thick till on the north west part of the basin and pass down almost to the basin floor. The presence of several chutes at the 365 m (1200 ft) level may reflect a relatively long stand of ice at that position.



Figure 4a. Stylized sequence representing former ice and proglacial lake positions during deglaciation. Dot patten is ice cover; horizonal lines are lake extents. See Figure 1 for location names and scale. This figure shows southward meltwater drainage phase.



Figure 4b. Eastward drainage controlled by basin topography.

The widespread occurrence of ice contact gravels, kettles and lacustrine deposits on the valley floor imply that the last ice had melted in the basin floor. However, several small upland valleys which lie as tributries to the present waterways also retained separated ice masses. In several, the flat floors contain hummocky topography and the walls have kame terraces. Leading out from these areas are meltwater channels which conducted water to either a lower ice surface or into proglacial lakes. In situations where the meltwater flowed into lakes small deltas formed. In other cases, fan deposits built at the base control. The presence of these deposits indicates that the downwasting stranded ice both large lowland and small upland basins. Thus the very last ice was not necessarily at the bottom of the lowest valleys.

The distribution and nature of surficial deposits is somewhat curious. First, compared to regions to the north (Lowell, 1980) the till is thicker. Not necessarily as a uniform blanket, but along depressions on the southern sides of hills, the till may be in excess of 20 m. The till thins as it laps onto the hill slopes and the hill tops expose bedrock. The presence of meltwater channels also has some relationship to the till distribution - the most common and best displayed channels occur lower on the hill slopes where the till is thick enough to be eroded.

Like the more abundant till, gravel deposits are more common in the Seboomook area than in the more northerly Allagash region. Gravel fans constitute the largest volume of stratified material in the area. Of course, abundance is relative; compared to the Allagash region the stratified deposits are large. However, when compared to the marine deltas, eskers, and outwash plains of southern Maine, they are insignificantly small. This impression may not be well conveyed, because the field trip visits the largest pits in the region.

This becomes more interesting when one considers that a fair measure of all the gravel probably originated from erosion as meltwater cut channels into the till. It seems then that the supply of englacial debris was small indeed in northwestern Maine - something one would expect near the center of an outflowing ice mass, but not from the outer portion of a large ice sheet.

LAKE BASIN HISTORY

A series of proglacial lakes occupied the large lowland basins throughout the Seboomook Lake area. The major basins discussed below include Seboomook Lake, (now dammed); Lobster Lake,(a natural lake); and Ragmuff Lake, (the name assigned to the basin now containing the West Branch of the Penobscot River). During deglaciation, the large basins commonly contained first stagnating ice masses and later proglacial lakes which exhibited falling lake levels throughout their history.

In contrast, small upland basins located within tributaries to the large valleys, show no evidence of occupation by former lakes. Instead these exhibit kame terraces along the valley sides and thin outwash deposits blanketing the valley bottoms. Some of these deposits prograde down valley, often via cols, and end in fan and deltaic deposits along the floors of the large basins. The field trip will examine the deposits in two of the large basins formerly containing glacial lakes.

Glacial Lakes Seboomook I & II

Damming of the West Branch of the Penobscot River created the present Seboomook Lake. This basin is relatively narrow and steep sided in comparison to other basins in the region and gives evidence for a complex history of ice down-wasting: final stagnation took place in the valley bottom and two different lake levels are present. The highest level (Glacial Lake Seboomook I) is evidenced by a delta with a topset-foreset contact of 340 m (1114 ft) a.s.l. and rhythmite sediments deposited to 336 m (1102 ft) a.s.l. (Stop 2). A second lower lake level, Glacial Lake Seboomook II at 331-333 m (1085-1091 ft) is indicated by two flat landforms (one exhibiting a topset-foreset contact) built into the lake. The lowest deltaic deposition may be associated with the valley fill in the North Branch of the Penobscot River (Stop 4) based on a calculated gradient of 2.7 m/km between the tops of the two features. All deltaic landforms are kettled and associated with hummocky topography, suggesting an ice-contact environment for both these lakes.

The exact drainage pathway for either of these lakes is unknown. Several possibilities exist and the two lakes may have shared the same outlet or had different outlets. Two channels - at 327 m (1071 ft) and 315+ m (1020+ ft) a.s.l. - exist between Seboomook Lake and Moosehead Lake to the south. However, marginal channels and kame terraces suggest that a stagnant ice mass existed in the Moosehead Lake basin and the eastern portion of the Seboomook basin at approximately 336 m (1100 ft), effectively blocking drainage to the south (Fig. 4c). Another alternative is that the water could have drained eastward through the same channel that the West Branch of the Penobscot River utilizes today. Lastly, the water may have drained eastward via a channel that leads from the Seboomook basin parallel to the modern



Figure 4c. Glacial Lake Seboomook I & II and Glacial Lake Lobster. Lake levels near 340 m. Relationship of Glacial Lake Seboomook and Glacial Lake Lobster not known.



Figure 4d. Glacial Lake Seboomook I & II (340 m) and Glacial Lake Ragmuff I (336 m). Possible drain for Glacial Lake Seboomook and Glacial Lake Ragmuff via paths represented by arrows.

Penobscot River. This channel begins at 336 m just above the east end of the lake and terminates at approximately 320 m. a.s.l. (Fig. 4d). If this channel was used as a spillway it indicates that the Seboomook Lakes would have drained into Glacial Lake Ragmuff I, which is lower than either Seboomook glacial lake.

Glacial Lake Lobster

Three successive lakes occupied a broad gently sloping basin which presently contains Lobster Lake, Little Lobster Lake, and the West Branch of the Penobscot River between Seboomook Lake and Chesuncook Lake (Fig. 1). Outlets from this basin include two low passes into the Moosehead Lake basin at approximately 314 m (1030 ft) and 320 m (1050 ft) as well as the modern valley extending towards Chesuncook Lake.

The highest lake in this series is designated Glacial Lake Lobster. This proglacial lake was confined to the basin of the West Branch of the Penobscot River as far north as Smith Halfway House, and extended into the basins of Lobster Lake and Little Lobster Lake if these were ice free (Fig. 4c). The limited evidence for this lake consists of rippled and rhythmically bedded sands found approximately 336 m (1110 ft) a.s.l. on the hill slopes of this basin. Ponded water at this elevation requires dams both to the south and to the north. The previously discussed ice block in Moosehead Lake, as well as an ice mass in the Chesuncook Lake basin, is hypothesized to provide such a base level control. Smith Halfway House was chosen as the northern margin of Glacial Lake Lobster based on field evidence of eroded ice-contact deposits at this natural constriction in the basin and because no evidence of this high lake level was found north of that location (Fig. 4c).

Whether Glacial Lake Lobster was continuous westward into the Seboomook Lake basin is problematic. Both Glacial Lakes Seboomook I and II existed at approximately the same elevation as this lake. However, no continuous shorelines or deposits can be traced from one basin to the other. The previously described channel between the two basins, as well as mapped deposits of till and ice contact stratified drift along the narrows between the basins suggest that a stagnant ice block may have separated the two lakes (Fig. 4c).

Glacial Lake Ragmuff I

Several pieces of evidence delimit the margins of the next lower lake, Glacial Lake Ragmuff I, in this basin. Shoreline features including lag gravels and washed bedrock knobs, as well as various exposures of rhythmically bedded sands, suggest a water level of 320 m (1050 ft) in a basin extending north to at least Chesuncook Lake. This lake also requires a dam at both its north and south ends. This could be accomplished by ice in both the Moosehead and Chesuncook basins or by higher lake levels in

both basins or by some combination of the two.

To the south, such a lake could have been continuous with the Moosehead Lake basin through three different passes: west from Lobster Lake, and south from the Penobscot River via two lowlands (Fig. 4d). A flat landform composed of pebble gravel at approximately 320 m (1050 ft) a.s.l. is found along the northeast shore of Moosehead Lake and traces into outwash up a tributary valley. This may represent the extension of Glacial Lake Ragmuff I into the Moosehead Lake basin.

To the west, Glacial Lake Ragmuff I was separated from the Seboomook basin because present topography rises above this lake level. The previously discussed channel feature originating at 336 m in the Seboomook basin and dropping to 320 m in the Lobster basin may imply contemporaneity of Glacial Lake Seboomook with Glacial Lake Ragmuff I (Fig. 4d).

Glacial Lake Ragmuff II

Glacial Lake Ragmuff II, the lowest lake, has deltaic features at the mouth of Luther Brook (Stop 9) and along Pine Stream (Hanson, unpub.) which measure 309-310 (1013-1018 ft) a.s.l. (Figs. 1 and 4e). These suggest a continuous water body from Lobster Lake to Pine Stream Flowage. The highly kettled delta in Pine Stream Flowage suggests that stagnant ice lay in this arm of Glacial Lake Ragmuff II. Because ice is present at this time, it implies that the previous lake (Glacial Lake Ragmuff I) also formed against an ice margin in the Chesuncook Lake basin (Fig. 4d).

Glacial Lake Ragmuff II extended into Lobster Lake and Little Lobster Lake, but had dropped below the spillway into Moosehead Lake (Fig. 4e). Thus its base control shifted north to the Chesuncook basin where it must have been dammed by ice or become continuous with a higher level Chesuncook Lake.

Fan Deposition and Riverine Terraces

Located at the mouths of several tributary valleys are fan deposits built into the glacial lakes. These tributary valleys contain a thin veneer of gravel along their floors. The gravels commonly extend down valley (often through cols) and end in fans and/or deltas. For example, evidence from Luther Brook (Stops 8 and 9) suggests that the fans were deposited into Glacial Lake Ragmuff II and at the same time eroded rhythmites previously deposited by higher lake levels. At Luther Brook the deposits shoe shallow deltaic stratigraphy, but more usually the fans exhibit braided channel, sieve, and plug deposits as well as imbricated cobble gravels. Throughout the field area, the indicators in these landforms consistently show flow into the large basins (rather than along their sides) and individual landforms begin at different elevations along the valley sides.

As such, these landforms have been mapped as fan deposits and constitute a major portion of the gravel within the field area. The distribution of fans suggests that some upland basins contained the last ice remnants.

Subsequent to fan deposition the rivers in the bottoms of the large lowland basins reworked the fan gravels into alluvial terraces. This interpretation stems from the observation that the lowest lying gravel deposits in the basins occur only in association with tributary valleys; rhythmically bedded lacustrine sands blanket the large areas between tributary valleys.



Figure 4e. Glacial Lake Ragmuff II (309 m). Location of deltas shown.

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ROAD LOG

Mileage	Description
0.0	Leave parking lot at south side of Seboomook Dam. Take first right (W).
0.3	Park along road, stop is on lake shore.

STOP 1: SEBOOMOOK DAM STRIATED OUTCROP

This outcrop displays cyclicly-layered sandstones and slate of the Seboomook Formation (Boucot, 1969). The sandstone beds show good examples of graded bedding. The blue-gray slate, widespread through out the area weathers to light gray after a short period of subaerial exposure.

The bedrock has been molded into whaleback forms by ice moving toward the ESE (122°). Superposed on the decimeter sized grooves are striations showing ice-flow towards the SE (154°). Although a 22° shift in ice-flow direction may not seem like much, it is suggested to be important because of its areal extent. The ESE flow occurs across northern Maine but the SE flow, present throughout the rest of the state, can not be traced much further north than this area.

How does this flow relate to the glaciation of the region ? If a southeastward flood of Late Wisconsin Laurentide ice is used to explain it, then why is it not present further to the north ? Alternately if the first Late Wisconsin ice-flow is to the ESE then the SE flow may relate only to ice covering Maine, not Laurentide ice. This problem requires discussion.

0.5	Pass Seboomook Dam campsite.
1.5	Keep left (S).
2.2	Turn right at access to Seboomook store.
5.7	Seboomook / Boyd east townline.

10.5 Pass 7 Mile Hill campsite.

11.6 Keep right to cross Beaver Brook (left goes toward Greenville). The flat area to the north consists of pebble gravel and may be a delta or outwash at 330 m (1085 ft) a.s.l.

12.7 Boyd east / Boyd west line.

14.1 Keep right at road to Greenville. Turn right (N) and take western road. Drive to and through campsite. Stop along lake shore.

STOP 2: RHYTHMITE SECTION AT SEBOOMOOK LAKE CAMPGROUND

This stop is located along the south shore on the southwestern corner of Seboomook Lake. The section is contains 7-9 m of massive and laminated sands and silts. It extends laterally 0.1 km towards the northwest where it is interbedded with or overlaps a mound of cobble gravel which may be an esker core (see Fig. 5).

The 8 m wide section to be examined on this trip is located on the west side of the larger section, and consists of three major units. The lowest unit consists of interbedded gray and tan sands with soft sediment deformation, loading, and faulting exhibited along a dipping upper contact. Rhythmically bedded sands and silts overlie the lowest unit which contains ripples, graded beds, and silt drapes. A maximum of 146 rhythmites have been counted in an adjacent section. The uppermost unit consists of massive tan sands. The unconsolidated portion of the section overlies pillow basalts of the Canada Falls Volcanic Member (Marvinney, 1983) which have well shaped ice-flow indicator showing former ice movement toward 142°.

The rhythmites are interpreted as lake bottom deposits. Underflows are responsible for the rippled and graded beds whereas deposition from suspension produced the sandy beds and silty drapes. The massive sand located at the top of the section is interpreted as shallow water lacustrine deposition in which the laminae have been disturbed by wind-generated turbulence, seiches, burrowing organisms (Ludlam, 1979) and/or convection of the epilimnion (Kempe and Degens, 1979). An added possibility to explain the lamination disturbance is mixing by frost and tree root action subsequent to lowering of the lake level.

This section crops out 0.1 km from a Gilbert-type delta composed of 1 m trough-cross bedded cobble gravel overlying 6.5 m dipping pebble sand beds. These beds dip at 11-20° and indicate water flow towards the east (78°) and northeast (47°). The topset-foreset contact of the delta measures 340 m (1114 ft) a.s.l. and indicates the water level at the time the delta was deposited. The top of the delta contains kettles, suggesting it was deposited over stagnant ice in the basin. The coarse sands exposed at the base of this section possibly formed in an ice contact environment prior to the formation of the rhythmite beds.

Bedrock knobs abut both sides and underlay this

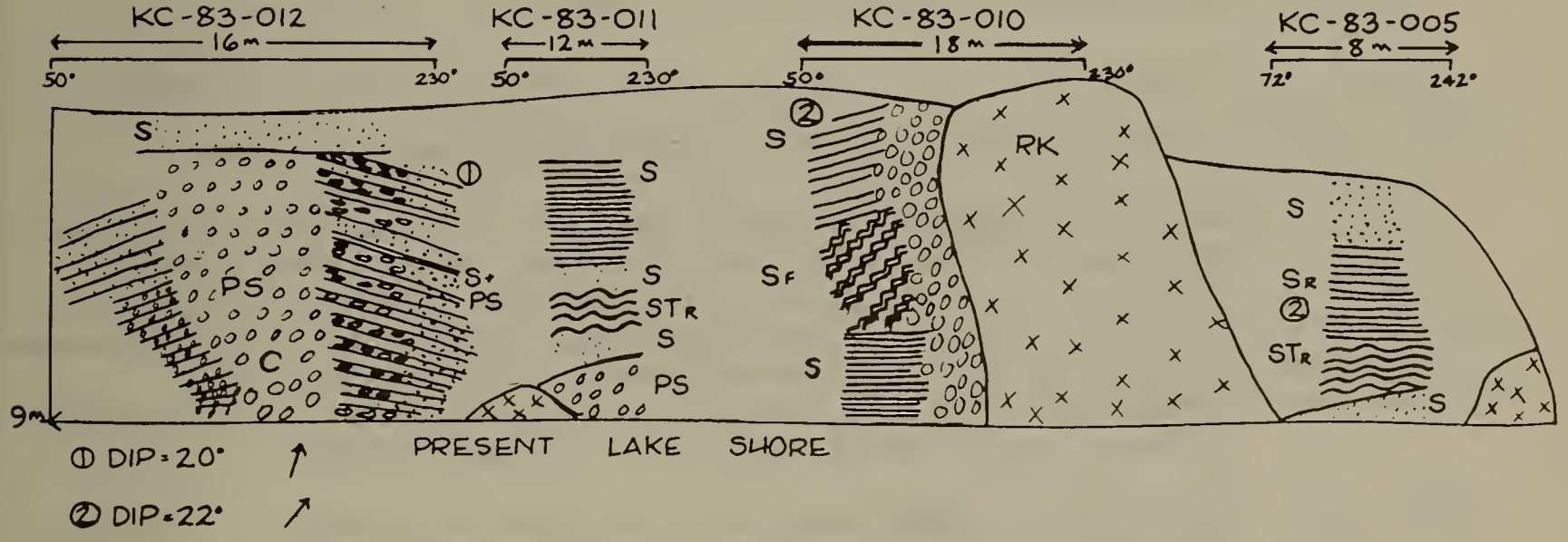


Figure 5. Rhythmite section of Stop 2. Stop will examine extream right hand side of exposure (KC-83-005). Legend for this and following figures: c/cg = cobble gravel, ps = pebble sand, = rippled beds, rk = bedrock, s = sand, st = silt. Circles numbers refer to current measurement locaties; arrow represents azmiuth of water flow. Scale and bearing of sections along edge of drawings.

rhythmite section. Bedrock exposures are a common feature all along the shoreline of Seboomook Lake, but only in the shallow western basin of the lake are the intervening areas between the knobs filled by lacustrine and ice contact sands and gravels.

14.6 Return to main road, turn right (W) to cross causeway.

15.4 Pass entrance to BSA Wilderness Base.

15.6 Cross South Branch of the Penobscot River.

15.9 Exposure in delta. Here 1.5 m of massive cobble gravel overlies 2.7 m of sand which dips 25° towards the east (113°). The topset-foreset contact is 333 m (1091 ft) a.s.l. and indicates a lower lake level than the rhythmites and delta at Stop 2.

16.0 Keep right (NW) at road to Canada Falls.

17.4 Pass Lane Brook campsite.

17.7 Turn right (N) at sign for North Branch.

19.3 Make hard left (W) toward Comstock Operation.

19.5 Turn right (W) into pit just beyond Comstock Camp.

STOP 3: PERIGLACIAL FEATURES AT COMSTOCK CAMP

This pit exposes massive and cross-bedded pebble gravel and sand. The west wall exposes imbricated gravel which indicates water flow towards the southwest (225°). The east wall exposes massive and cross-bedded pebble gravel with sand lenses. Dip directions indicate flow towards the southwest (205°). A braided stream environment deposited these gravels.

On the northeast wall 5 m of massive pebble gravels have kink folds and vertically aligned clasts that cut the horizontal beds. Compression from an enlarging ice wedge and summer expansion against it may explain the folding of the pebble beds adjacent to the vertical pebbles. Slump and subsequent melting filled the wedge with vertical pebbles. Thus cryoturbation and fossil ice wedge casts would indicate a former permafrost environment here. Periglacial features have been recorded for other scattered locations in northern Maine (Lowell, 1980; Newman, pers. comm., 1981; Kite, pers. comm., 1982).

Turn around, head north past Comstock camp.

20.2 Cross bridge at Leadbetter Falls. Note horizontal bedding in pit exposure.

20.4 Turn left (W) and drive along top of pit.

20.7 Turn left (W) into pit.

STOP 4: TERRACE DEPOSITS AT LEADBETTER FALLS

Two pits are exposed in this deposit which constitutes the lowest glacial landform in the valley of the North Branch of the Penobscot River. The landform abuts the valley wall on the east and has been eroded by the river to the west (a terrace step is eroded into the landform along the west side of the northwestern pit).

The northwest pit (this stop) contains 12.5 m of massive, aligned, and cross-bedded cobble gravel and pebble sand (see Fig. 6). Dipping beds indicate water flow towards the south (174°) and southwest (228°), while ripples within sandy lenses indicate flow towards the southeast (131°) and west (282°). Flow parallels the modern valley. High angle faults, a filled kettle, and the steep ice contact face of the pit indicate outwash deposition over and adjacent to ice. The large scarp on the northwest portion of the landform is due, in part to erosion by a meltwater channel from the northeast.

The southeast pit, viewed from the road, exposes 6-7 m of imbricated and cross-bedded cobble gravel and pebble sand. Flow directions (based on imbrications and apparent dip) vary from northeast to southeast (50°, 90°, 115°, and 132°), but generally indicate flow parallel to the modern valley. An outwash origin for this deposit is suggested. The boulders measured in the two pits in this landform show a decrease in size as distance increases away from the ice contact portion of the deposit.

The top of the pit measures 348 m (1141 ft) a.s.l. and indicates a valley fill estimated to be 15 m (50 ft) thick. This feature is unique in the region and suggests that valley trains did not commonly fill the valleys and later erode to form valley terraces.

21.0 Return to main road, turn left (N).

21.6 T4 R18 / T4 R17 town line.

21.8 Turn left (NW) at stop sign onto the Golden Road.

22.0 T4 R17 / T4 R18 town line. Note numerous pits along roadway.

25.3 Pull into pit with canvas shed.

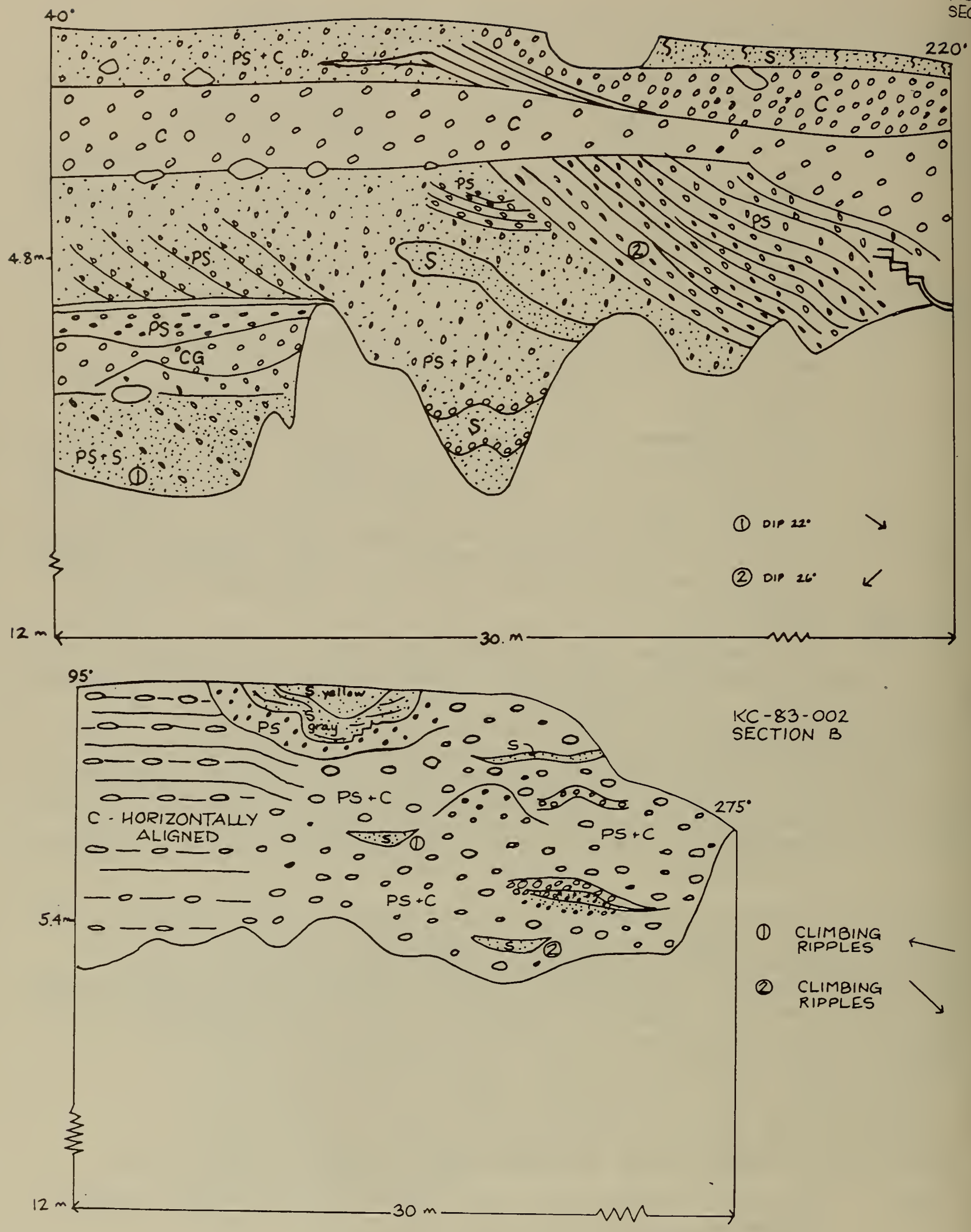


Figure 6. Pit exposure at Stop 4. Legend same as Fig. 5.

STOP 5: FAN DEPOSIT AT SPENCER BROOK

This 10-12 m thick deposit sits above the east side of the North Branch of the Penobscot River. Massive cobble gravels and pebble sands are interbedded with rippled and cross-bedded sands and pebble sands. Dipping beds and ripples indicate flow towards the southwest (225°) ie. into the present valley. A small kettle can been seen near the center of the pit. This gravel deposit overlies till exposed lower in the valley wall approximately 0.25 km southeast of this pit.

The landform is interpreted as a fan deposited by Spencer Brook. This fan lies at the same elevation, 366 m (1200 ft), as another similar deposit 3.5 km further north up the valley. This second smaller fan lies at the mouth of Bog Brook, but both fans appear to emanate from the same upland basin and may have been fed from a stagnant ice mass melting there.

A third 366 m (1200 ft) landform occupies the center of the valley north of Bog Brook and has an ice-contact face along its north side. The consistent elevations of the three landforms suggest a deltaic origin, but no clear deltaic stratigraphy is exposed in any of the sections. Therefore, these deposits are interpreted as fans built into standing water as described by Axelsson (1967). Plugs of boulders, imbricated cobbles, channel deposits and sieve deposits present within described sections through out the valley substantiate the fan hypothesis (Reading, 1978).

Such fan features are common to all basins in the field area and several have been mapped along the north shore of Seboomook Lake. However they are thin and cover the hillside with a blanket morphology. The interpretation of these features as fans is based on the following evidence: 1) the gravel deposits become thinner towards the north (upslope), 2) the grain size of the deposits coarsen towards the north (upslope), 3) the apices of different deposits begin at different elevations above the lake, 4) the flow indicators consistently show flow into the lake basin, and 5) these features cannot be traced into a flat continuous terrace that parallels the lake shore. Additionally, the fans are commonly fed from cols containing channels that end upstream in small basins where ice blocks could have produced the meltwater to feed the fans.

26.2 Pass good exposure of bedrock on the right and point bars on the left.

27.3 Turn left (NW) at Snake campsite.

LUNCH STOP: SNAKE CAMPGROUND ON THE NORTH BRANCH OF THE PENOBSCOT RIVER

Note the imbricated gravels in the stream bank and the silty alluvium which overlies them in a filled meander scar north of the picnic area.

28.2 Return to the Golden Road, turn right (SE).

33.7 Return to junction at mile 77, continue east.

34.5 Turn right onto "Com 1".

35.1 Park at small turnout on left (E) side of road.

STOP 6: CHANNELS ON NORTH SHORE OF SEBOOMOOK LAKE

The road runs along an interfluve between two submarginal channels that are cut into thick till. The incision is approximately 30 m; the channels begin near 410 m and end at the 350 m level. This is one of four channels on the northwest side of Seboomook Lake. At their upper end they parallel contours for a few hundred meters before they dive downslope in a zig-zag pattern for distances up to 2.5 km.

The channels may emanate from downwasting ice which fills the basin to the 410 m level. Meltwater freely drains towards the bottom of the ice where it intersects a base control (piezometric surface). These features represent a late phase of deglaciation because any englacial meltwater flow above the 365 m level could pass through high cols out of the basin in a southerly direction. However, once the base level drops below this level it must move eastward paralleling modern drainage, through the basin.

One interesting aspect of the submarginal channels in this area is that a smaller submarginal channel obliquely crosses the tops of the larger channels. The relative age of these features is problematic.

35.7 Return to the Golden Road, turn right (E).

36.1 Turn right (S) onto "Seboomook Loop Road".

38.3 Road drops from thick till onto channel floor.

39.3 Turn right (S) onto unmarked road. Note topography.

39.7 Note rock scarp and Pittston Farm to right (SW).

40.0 Park at end of road.

STOP 7: HUMMOCKY TOPOGRAPHY ON THE NORTH SHORE OF SEBOOMOOK LAKE

Depositional hummocky topography consisting of gravel hummocks and ridges with kettles comprise the landscape here. These drape over bedrock knobs which also show fair relief. This topography occurs ubiquitously in the western portion of Seboomook Lake basin. Although some ice contact gravels extend to levels of 365 m (1200 ft) and higher in some valleys, the major concentration is at 335 m (1100 ft). a.s.l. These deposits lie directly across the lake from the rhythmite section (336 m a.s.l) examined at Stop 2 and the kettled delta (341 m a.s.l) adjacent to it.

This suggests a period in the history of this basin when sediment was being deposited over stagnant ice blocks and building a delta into the basin. Contemporaneous with or slightly following this event, sandy and silty rhythmites were deposited into the same area of the lake basin while the water level remained at an elevation of at least 336 m (1102 ft) a.s.l.

40.7 Return to Seboomook Loop Road, turn right (E).

41.0 Note pit in ice contact drift on left (N) side.

41.6 Note Seboomook Lake on right (S).

41.9 Pittston / Boyd west town line.

43.2 Return to the Golden Road, turn right (E).

46.1 Boyd west / Boyd east line.

49.7 Cross Nulhedus Stream.

51.9 Note 4 way intersection (southern road leads to Seboomook Dam).

55.2 Pass road north to St. John Operation.

57.3 Seboomook / Burbank town line.

57.4 Turn right (S) onto unmarked road just before rock crop.

57.6 First part of stop (A) is exposure on right (W). Remainder of stop (B) is in this pit further to the west.

STOPS 8 & 9: LACUSTRINE SEDIMENTATION AND EROSION AT LUTHER BROOK

These last two stops examine three separate exposures in a single landform at the mouth of Luther Brook, a

southward draining tributary to the West Branch of the Penobscot River. The stratigraphy of the two stops will be described separately, followed by an interpretation which includes both stops.

STOP 8

The first section (east end of pit) consists of 1.77 m of rhythmically bedded sands and silts. The laminae vary in thickness from 20 to 85 mm and include ripples, drapes, and graded beds. The base of the section shows thicker sandy beds with silt drapes, while thinner silty rhythmites overlie them. The uppermost beds are mottled silt devoid of structure. Prior to pit excavation, a thin cobble gravel may have been the top unit (see Fig. 7).

The second section (west end of pit) is composed of 30 cm pebble sand which contains cobbles and which overlies 1.5 - 2 m sand beds dipping towards the south (162°). The contact is 309 m (1013 ft) a.s.l. Below the dipping beds are sands containing assorted pebbles and clasts of silt, sand, and gravel which have been interpreted as rip-up clasts. This lowest unit underlies the foresets in the west wall of the pit but is best exposed in the gullies on the south side of the pit.

58.2 Return to the Golden Road, turn left (W).

59.0 Pass road 19A to north.

59.2 Turn left onto unmarked road.

59.5 Park at corner.

STOP 9

This section contains 3.2 m of cross-bedded pebble sand and gravel overlying rhythmically bedded sands and silts. The rhythmites vary in thickness from 50 to 95 mm, and are interpreted as lake bottom deposits. The pebble sand, which includes cobble gravel, overlies an erosional contact which truncates the horizontal laminae of the rhythmites (see Fig. 8a). The truncated surface is 306 m (1002 ft) a.s.l and the top of the gravel section is 309 m (1012 ft) a.s.l.

The rhythmites are commonly sandy beds with silty drapes, graded beds, ripples, and soft sediment deformation. One lens of coarse gravel is included in the rhythmites. These beds vary laterally to a silt/clay pod. An soil auger hole shows the pod to be 66 cm thick and overlies at least 10 cm of coarse sand

KC-83-069

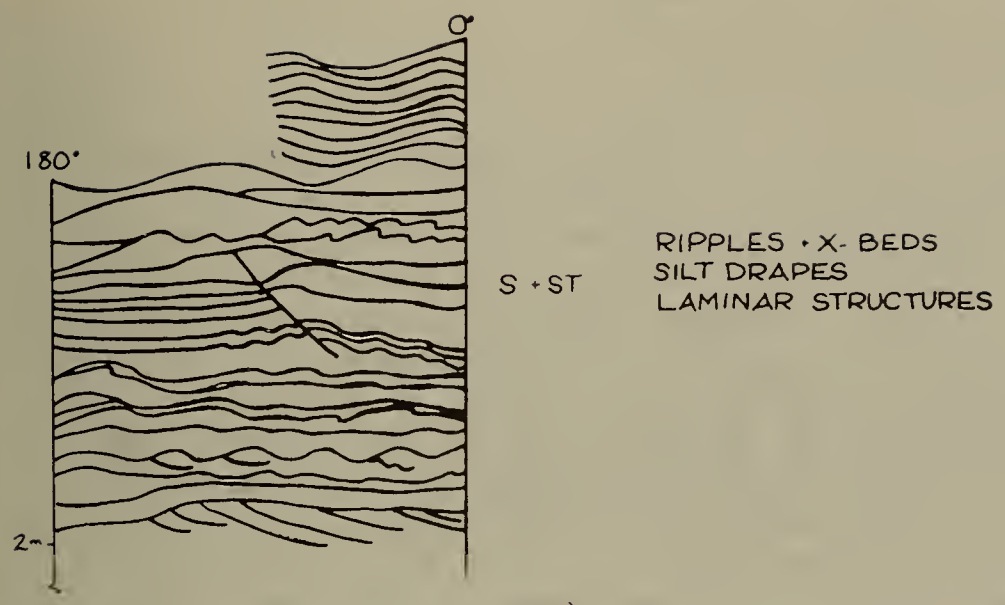


Figure 7. Exposure at Stop 8a.

Lake bottom sediments.
Legend same as Fig. 5.

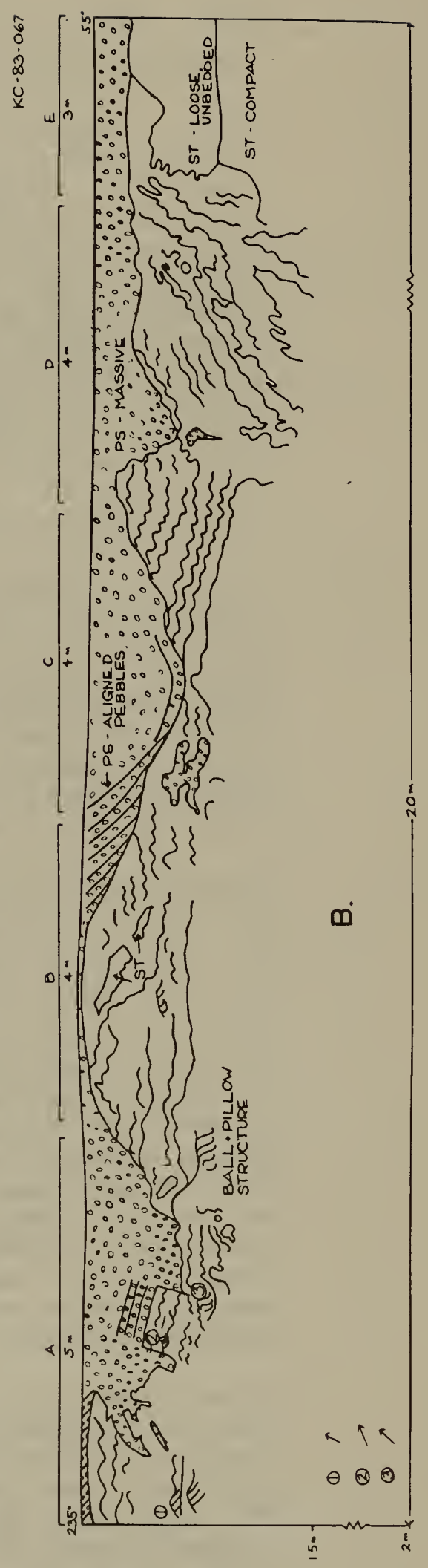


Figure 8a,b. Exposures at Stop 9. Legend same as Fig 5.

with angular pebbles (see Fig. 8b).

INTERPRETATION OF STOPS 8 and 9.

Both of the above pits are located in a flat landform now incised by three streams, two of which originate from the Luther Brook valley. Three additional pits, located up Luther Brook valley from this landform and not visited on this trip, expose pebble gravel which extends to 328 m (1077 ft) a.s.l. and overlies till. However, unlike classic valley fill, this gravel is spread thinly over both the valley bottom and sides. No terraces cut by the modern stream interrupt the smooth valley floor. The gravel exposed in the upper 3 pits appears continuous with the back of this flat landform containing the lower pits. This suggests that the Luther Brook valley was the source for the gravel producing the flat landform.

The stratigraphy exposed in the entire landform suggests the following history. First ice blocks sitting in the Luther Brook valley deposited the high gravel. These ice blocks may have been contemporaneous with the highest level (335 m) of Glacial Lake Lobster in the Penobscot Valley. Rhythmically bedded sands and silts (examined at stops 8 & 9) were deposited into Glacial Lake Lobster and a subsequent lower lake which measured 320 m a.s.l. (Glacial Lake Ragmuff I).

When the lake dropped to an elevation of 309 m (1020 ft), this flat landform built into Glacial Lake Ragmuff II. The following evidence suggests a erosional event in the Luther Creek basin: 1) the erosional contact on the top of the rhythmite section, 2) the lateral discontinuity of different rhythmite sections, 3) the rip-up clasts which underlie the foresets, 4) the limited thickness of foreset beds, and 5) the restricted extent of the deltaic deposits. We see the above evidence as indicating a flood event which eroded the former lake bottom in a local area and filled the eroded portions with sediment containing rip-up clasts then built a delta into shallow water (Stop 8, west section). In nearby areas where the erosion was not as great, gravel was deposited over or adjacent to eroded lacustrine beds (Stop 8, east section, and Stop 9).

This lowering of the lake level also eroded portions of the ice contact gravels in the Luther Brook valley and left only a thin cover of gravel over till at the apex of the gravel deposit. The entire landform including the high thin gravels, and both the lower erosional and depositional facies is mapped as a fan deposit with Luther Brook provided the water for the erosion of the rhythmite beds and the deposition of the deltaic sands and fan gravels.

59.9 Return to the Golden Road, turn left (W).

63.8 Pass mile 65 sign.

64.3 Turn left (S) toward Seboomook Dam.

67.2 Cross Seboomook Dam. End of Log. Return to Greenville by retracing first portion of road log to mile 11.6, turn left (S), continue straight through all intersections; about 20 miles to Rockwood. Turn left after crossing the Moose River, follow Rt. 6 and 15 to Greenville (1:15 total travel time).

TRIP C-5

OPHIOLITE AND MELANGE TERRAIN CAUCOMGOMOC LAKE AREA, NORTHWESTERN MAINE

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INTRODUCTION

Outcrops of Pre-Silurian rocks in the Caucomgomoc Lake area of northwest central Maine are interpreted as a stratigraphic succession of ophiolite overlain by metasiltstone melange. The rocks are greenschist metamorphic grade and have an estimated minimum cumulative stratigraphic thickness of 4000 m. The Pre-Silurian section of the Caucomgomoc Lake area occurs as an isolated structural fenster in fault contact with upper Silurian (Pridoli) to lower Devonian (Gedinnian?) sedimentary and volcanic rocks on the east, south and north. Locally to the south and north, the Pre-Silurian section is in fault contact with the Seboomook Formation (lower Devonian - Seigenian). The faults are not extensively exposed. However, where they are exposed, they show pronounced cataclasis and mineralization to include quartz, limonite, and calcite. Most of the major faults appear as high angle easterly dipping normal faults, with the hanging wall down dropped to the east. Over all, the Pre-Silurian dips and youngs to the west. On the western outcrop margin the melange is unconformably overlain by the Frontenac Formation.

The area had not been mapped in detail prior to this study. There were geologic reconnaissance excursions in the 19th century by Jackson (1838) and Hitchcock (1861, 1901) and in the 20th century by Raabe (1973). Over the last decade, the ophiolite has been investigated by private industry and consultants for economic mineral deposits. Geologic maps of this area include Hitchcock (1901), Keith (1933), Doyle (1967) and Raabe (1973). These show the area to consist variably of sandstones, slates and greenstones (metabasalts). Pollock (1983, in preparation) is the latest contributor to the geology of this area.

STRATIGRAPHY[1]

The Pre-Silurian section is composed of three formations which generally thin to the north. The lower most unit, (COl

[1]The Geologic Names Committee of the U. S. Geological Survey has approved the use of following formation names: Loon Stream Formation, Caucomgomoc Lake Formation and Hurd Mountain Formation. These are being used informally here with the committee's permission. Formal descriptions are in preparation.

on figure 1) has not been subdivided into members of lentils. The middle formation (COc on figure 1) has three members in the vicinity of Caucomgomoc Lake. These members undergo a facies change to the north and south and/or are tectonically terminated against basalts of this formation to the south and north, and metasiltstone of the uppermost formation to the north. While the upper unit (COh on figure 1) is interpreted as a melange. Three separate lithogically distinctive members are recognized in this formation.

FORMATIONS

Unit 1. (Loon Stream Formation)

The lowermost formation is a very fine to fine grained pale green (10G 6/2), grayish green (5G 6/1) and pale yellowish green (10GY 7/2) metapelite. Locally red colorations may be seen. The metapelite is characteristically slate and phyllite. Less than 1% detrial grains or silt size grains are seen in thin section. Subsidiary lithologies within this formation include medium to coarse grained aphyric metabasalt, phyric metafelsite with phenocrysts of biotite and quartz and poorly sorted metawacke.

Sedimentary structures are poorly defined or preserved in the metapelite. Where bedding is clearly seen it is 2 to 20 cm thick and is delineated by ubiquitous thin white to buff weathering laminae. The laminae are interpreted to have originally been silts and/or volcanic ash. The laminae are generally structureless, but rarely exhibit textural grading. Other sedimentary structures have not been observed.

This unit does not crop out extensively. Outcrops most commonly do not exhibit sedimentary or tectonic structures other than a pervasive slaty cleavage or schistoscity which trends northeast and dips variably to the northwest. Because of a lack of sedimentary structures, it is difficult to estimate preserved stratigraphic thickness. Thickness estimates range between 1000 and 3000 meters.

The contact zone between this and the overlying unit varies locally. The contact is either conformable and gradational or unconformable. In both instances wacke beds increase in abundance and thickness upward toward the contact with the overlying unit. On the south shore of Caucomgomoc Lake there occurs a zone where wacke beds and metapelite are disrupted and are not physically traceable for distances of more than a few meters. This disrupted zone is clearly evident along the shoreline south of an unnamed island 400 meters off shore. On the north shore of Caucomgomoc Lake the transition between the two units appears gradational and conformable. An interpretation given to account for this variation is that deposition of the lower unit occurred on a slope, or that the sediments were semi or unconsolidated. Volcanism

responsible for the rocks of the overlying unit caused the slope or sediment to become unstable, initiating slumping or other mass movement. The result was local production of a disrupted zone between the lower and overlying units. Another interpretation assumes that the whole Pre-Silurian package is a tectonic melange, into which is incorporated very large sections or rafts of ophiolite. Within this package are tectonic zones or imbricate thrusts which are local and are not internally persistent for large distances.

Unit 2. (Caucomgomoc Lake Formation)

This unit consists of a succession of interbedded volcanic and clastic sedimentary material which is 1500 to 2400 meters thick. Generally the unit is thicker in the southern part of the outcrop area. In the vicinity of Caucomgomoc Lake the unit is subdivided into three members. North and south of the lake rapid complex facies changes and poor outcrop make subdivision of the unit impractical. Where mapped, the lower and upper members are volcanic in origin and the middle member is sedimentary. The rocks young westward with dips ranging between 37° and 90° to the west. Small synclines and anticlines are present on the north shore of Caucomgomoc Lake. The section to the south of Caucomgomoc Lake is essentially a monocline with strikes between N 10° W and N 35° E. Dips average 57° to the west.

The volcanic rocks are basaltic in nature. Usually SiO_2 content ranges between 44 and 51 wt%. Alkali content ($K_2O + Na_2O$) exceeds 3.5% in most samples. TiO_2 commonly exceeds 1.5%. The rocks have been metamorphosed to greenschist facies. For the most part the original minerals have been recrystallized and textures include ghosts or relics of pre-existing minerals. The mineralogy of the volcanics is chlorite, epidote, zoisite, calcite, ± amphibole, ± biotite, ± plagioclase (albite or oligoclase), ± quartz, ± Fe-Ti oxides and sulfides. Relic pyroxenes are only very rarely present and olivine has yet to be observed. These rocks would best be referred to as spilites following the usage of Williams, Turner and Gilbert (1982). Details on the petrology is in preparation.

Lower Member

Thickness of this member ranges between a maximum of 1100 meters south of Caucomgomoc Lake to 196 meters and less north of the lake. The member is comprised of metabasalts. Specifically these metabasalts are agglomerates, breccias, lapilli tuffs, tuffs and flows. Metapelite, chert, and wacke (quartz and felspathic) are rare lithologies interbedded with the agglomerates, lapilli tuffs and flows. This member differs from the upper member in that it is principally comprised of a larger volume of agglomerates and lapilli tuffs relative to flows. The upper member is comprised of a larger volume of flows relative to agglomerates and lapilli tuffs. The agglomerates are composed of bombs ranging in size from 6.4 to 35 cm. The bombs are commonly aphyric

EXPLANATION

LOWER DEVONIAN

Dsu — SEBOOMOOK FORMATION UNDIFFERENTIATED

ORDOVICIAN(?) TO DEVONIAN(?)

ODf — FRONTENAC FORMATION

UPPER SILURIAN AND LOWER DEVONIAN

ALLAGASH LAKE FORMATION

DSav - predominantly pillowed baslat with interbedded conglomerate, limestone and sandstones of various classification

DSas - mixed sedimentary rocks including limestone, wackes, arenites, siltstones and shales (with and without hematite cement) and conglomerates.

CAMBRIAN(?) AND ORDOVICIAN(?)

UNIT 3 (Hurd Mountain Formation)

COhb - pillow basalts, basaltic agglomerate and basaltic lapilli tuffs

COhm, COhs - meta-siltstone, claystone slate and phyllite minor calcareous meta-siltstone. Commonly exhibits rusty weathering. Unit is complexly deformed and pervasively sheared. Unit is interpreted as a tectonic melange. Other rock types in the pelitic host include quartz wacke, meta-basalt, ultra-mafics, gabbros, diorites and granodiorites.

COhs - medium to thick bedded quartzose wacke, usually texturally uniform and lacking sedimentary structures.

UNIT 2 (Caucomgomoc Lake Formation)

COcb - a dominantly phyric and aphyric meta-basalt flows undifferentiated.

$COcb_2$ - upper member on Caucomgomoc Lake. Phyric and aphyric pillowed and non-pillowed basalt flows, basaltic agglomerate and basaltic lapilli tuff.

COcs - medium to thick bedded quartzose wacke and siltstone.

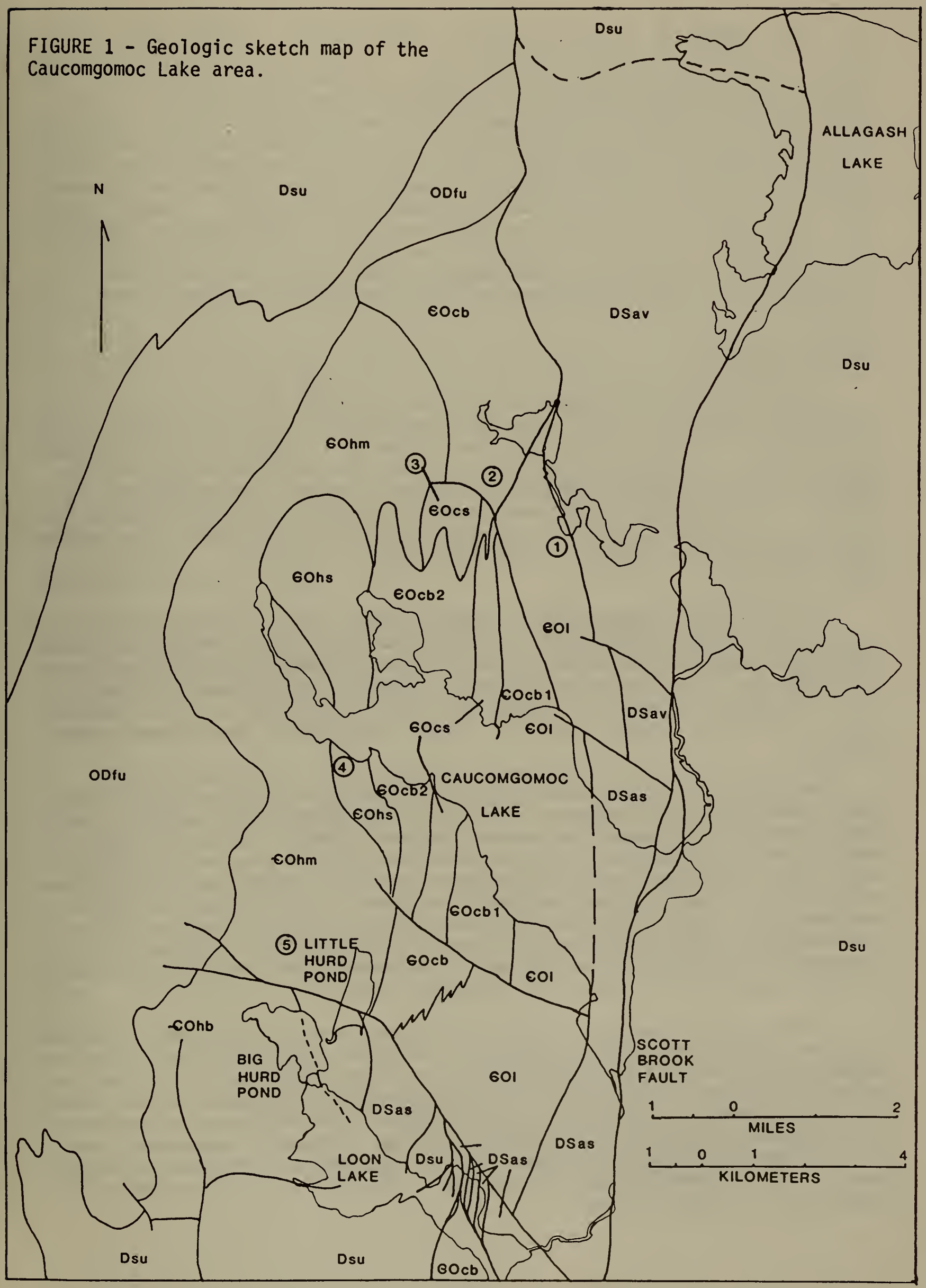
$COcb_1$ - lower member on Caucomgomoc Lake. Basaltic lapilli tuffs, basaltic agglomerates, basaltic flows common.

This unit is locally intruded by gabbroic and ultramafic stocks and dikes.

UNIT 1 (Loon Stream Formation)

CO1 - greenish meta-pelite including siltstone, slate and phyllite. Thin to thick bedded ubiquitous laminae and grayish red color common locally.

FIGURE 1 - Geologic sketch map of the Caucomgomoc Lake area.



and less commonly to rarely phyric metabasalt. Bombs less than 12 cm in size are usually ovoid or elliptical in nature, and appear flattened in the schistocity. Larger bombs exhibit a variety of streamlined shapes, are more commonly phyric or amygdaloidal and exhibit chill margins 1 cm or more thick. Matrix of the agglomerates is ash and lapilli. The lapilli are most commonly aphyric metabasalt. The flows in this member are commonly non-pillow structureless or amygdaloidal aphyric metabasalt. Individual flow thickness is difficult to determine. Bases of the flows where seen exhibit chill margins. The tops of the flows are irregular with flow lines in the upper several centimeters. Brecciated flow tops are uncommon. Locally, pillow structures 10 to 25 cm in size are observed in the upper portion of the flows.

Middle Member

Thickness of this member ranges from 511 m south of Caucomgomoc Lake to 215 m and less north of the lake. This member is predominantly composed of quartz wacke. Feldspathic or plagioclase arkosic wacke, metapelite, pebble to boulder conglomerate with a pelitic matrix, chert and metavolcanics (basalts) are less common. The wackes are commonly subrounded, composed of fine to coarse sand sized grains supported in a matrix of very fine grained white mica and chlorite. Calcite cement is common. Sorting is typically poor.

Sedimentary structures in the wackes vary. Bedding thickness ranges from approximately 30 cm to more than 3 m. Where exposed upper and lower bedding surfaces are relatively flat and uniform. The beds are usually structureless except for uncommon normal and/or reverse grading, poorly developed parallel or flat lamination, rare small scale current ripple lamination and mega-ripple bedding.

Depositional mechanism of the wackes is interpreted to have been from turbidity currents for thinner beds which exhibit partial or complete Bouma Sequences (Bouma, 1962). A grain flow mechanism is envisioned for thicker beds which are generally massive, but which may exhibit reverse grading, combinations of normal and reverse grading, and poorly developed parallel or flat laminations (Middleton and Hampton, 1973). Grain or fluid flow indicators such as dish structures or fluid escape pipes which may have originally been present have been obscured or destroyed by deformation processes and metamorphism.

The conglomerates consist of a shaly or pelitic matrix which enclose subrounded pebble to small boulder sized clasts or quartz wacke and metabasalt. Additionally, these contain discontinuous broken and folded wacke beds. Bedding thickness varies and upper and lower bedding planes are irregular. The clasts are randomly distributed throughout the shaly matrix. The conglomerates are interpreted to have originated as debris flows. These are uncommon and are interbedded with wackes interpreted to have been deposited as grain flows.

Upper Member

This member varies in thickness from approximately 760 m and less south of Caucomgomoc Lake to 1000 m and more north of the lake. This member is primarily composed of pillowed meta-basalt flows of variable (3 to 15+ m) thickness. Non-pillowed flows are also present and are generally thicker than the pillowed flows. Minor lithologies include basaltic agglomerates and lapilli tuffs, wackes, chert and pelite.

The flows are most commonly phyric with relic amphibole biotite or plagioclase phenocrysts, or amygdaloidal with amygdules of calcite, chlorite and epidote. Phenocrysts and amygdules are small, usually less than 4 mm. Thicker non-pillowed flows may be differentiated with concentrations of small relic mafic phenocrysts in the lower portion and relic plagioclase phenocrysts in upper portions. Pillows range in size from 6 cm to as much as 2 m. Average pillow size is less than 45 cm, with chill margins less than 3 cm thick.

Agglomerates, lapilli tuffs and sedimentary rocks which occur in this member are similar to those described previously in the lower and middle members.

Undifferentiated Units

Three undifferentiated sections occur south of Loon Lake, east of Little Hurd Pond and northwest of Poland Pond. These are metabasalts similar to those described for the upper member in the vicinity of Caucomgomoc Lake.

Intrusives

This formation is intruded by numerous dikes and by a layered meta-gabbro and meta-pyroxenite. Intrusives have been metamorphosed to greenschist facies. The dikes range in size from 6 cm to more than 1 m in width. Compositionally they range from felsic to ultra-mafic and are phyric or aphyric. Phyric varieties commonly are flow differentiated. The layered meta-gabbro and meta-pyroxenite crops out on the south shore of Caucomgomoc Lake with several low water exposures in the lake itself. Over all the outcrop is sub-circular. Originally, this may have been a major feeder pipe which contributed to the volcanic pile of this formation.

Unit 3. (Hurd Mountain Formation)

This unit consists of three principle lithologies or members. These are 1) meta-quartz wackes (COhs); 2) meta-siltstone and meta-claystone (COhm); 3) metabasalts (COhb). The contacts between these units and the underlying formation are not exposed but detailed mapping indicates this unit structurally overlies and is laterally juxtaposed to the preceding formation.

The quartz wacke (COhs) may be conformable on top of meta-basalt. There is reasonable consistency of strikes and dips across the contact area to support this supposition.
The quartz wackes are medium to thick bedded, texturally uniform and generally lack current generated sedimentary structures. This unit is thicker in the northern outcrop area than in the southern. It is not known if the thinning is due to an original sedimentary facies change or a later tectonically induced thinning. The latter is the preferred interpretation.

The second lithology or member is dominantly a laminated or thinly bedded meta-siltstone, claystone slate or phyllite. This lithology commonly may be characterized by rusty weathering.

Locally the meta-siltstone and claystone slate or phyllite is a pebble conglomerate. Pebble sized clasts of pelite or aphanitic igneous rocks and medium to coarse sand size grains of quartz and plagioclase feldspar are completely supported in matrix. Pebble sized clasts are commonly elongated and attenuated in the schistoscity. The pebble sized grains are not ovoid or elliptical but rather elongated angular grains with ragged or serrated ends. Minor lithologies within the meta-siltstone include calcareous siltstone, quartz wacke, metabasalt and ultramafics, primarily pyroxenite. Intruded (?) into this unit are small granodiorite and gabbro stocks and small discontinuous porphyritic aphanite sills.

The meta-siltstone lithology is interpreted as a tectonic melange. The thin bedded meta-siltstone is the host lithology. The other lithologies mentioned are tectonically included remnants of previously more extensive units or intrusions(?). The meta-siltstone is pervasively sheared. Thin discontinuous siltstone lamina or beds are tightly folded into kink bands or chevron folds. Thick wacke beds are also discontinuous, exhibit a variety of fold styles and are commonly broken by small scale faults. The metabasalts and ultramafics are of limited outcrop extent, ranging from a few square meters to several hundred square meters in area. The preferred interpretation is that the wackes (COhs) and the metabasalts (COcb) and wackes (COcs) were partially dissected during the subduction or obduction process.

The third lithology of this unit is a large relatively thick section of metabasalts, very similar in nature to the basalts on Caucomgomoc Lake. The section is predominatly pillowed and non-pillowed metabasalts. Minor lithologies include agglomerates and lapilli tuffs. The lower contact of this section is exposed on the shore of Big Hurd Pond. Structural relationships indicate that this overlies meta-siltstone. A conglomerate at the base of the section has pebble sized clasts attenuated and elongated into cleavage These basalts were incorporated into the melange during subduction. The northern and southern contacts are fault contacts either with the meta-siltstone melange or with the Seboomook formation.

STRATIGRAPHIC CORRELATION OF OPHIOLITE AND MELANGE MAINE AND QUEBEC

	QUEBEC Laurent, 1978	WESTERN MAINE Boone, Boudette and Moench, 1981 Boudette, 1982 Boudette and Boone, 1976	CAUCOMGOMOC LAKE Pollock (this paper)	MUNSUNGUN Anticlinorium Hall, 1969, 1970
LOWER ORDOVICIAN	St. Daniel Formation			
CAMBRIAN	Upper Ophiolite Complexes	Hurricane Mountain Formation	Unit 3 Hurd Mountain Formation	Unnamed Green Phyllite Chase Brook
		Jim Pond Formation	Unit 2 Caucomgomoc Lake Formation	
			Unit 1 Loon Stream Formation	
	Lower Ophiolite Complexes	Boil Mountain Formation		

Age and Correlation

The age of the units described here has not been established. Detailed mapping to date has not led to the recovery of macro fossils. Selected lithologies have been appropriately treated and examined for micro fossils without positive results. The entire section is currently thought to range in age from middle to late Cambrian through lower Ordovician(?). The inferences to the age are based on correlation with similar lithologies and stratigraphic sequences to the southwest and northeast of the Caucomgomoc Lake area within Maine and to the west in Quebec.

Figure 2 shows current stratigraphic correlations of the Caucomgomoc Lake ophiolite and melange with other sections in Maine and Quebec.

The section described in this paper is stratigraphically homotaxial with the Jim Pond and Hurricane Mountain Formations, (Boone, Boudette and Monech, 1981; Boudette, 1982; and Boudette and Boone, 1976) of western Maine. This section was presumed to be of late Cambrian to early Ordovician age in these articles. However, G. Boone (verbal communication, 1982) assigns a Cambrian age to the Hurricane Mountain Formation. Correlatives to the northeast include the Chase Brook Formation and an unnamed green phyllite. (Hall, 1970) and Group A unit 1 and 2, (Hall, 1969). Hall, (1969, 1970) assigns a middle and upper Cambrian age to these units. Laurent (1978) assigns an upper Cambrian to lower Ordovician age for ophiolites of the Thetford Mines area and lower Ordovician for the melange of the St. Daniel Formation in Quebec.

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ITINERARY

TRIP		
0.0	0.0	8:00 a.m. leave from Pray's store and travel west on the Golden Road. Note: This trip involves travel on remote privately maintained woods roads where access is controlled. Your vehicle should be equipped with good tires, a good spare and a reliable

		jack. You should also have a full tank of gas at the start. This trip would not have been possible without the permission and courtesy of North Maine Woods, Ashland, Maine; Seven Islands Land Company, Bangor, Maine; and Great Northern Paper Company, Millinocket, Maine. The itinerary lists several optional stops. One or more of these may be visited in lieu of those planned. Those outcrops visited will depend on weather, lake level and road conditions on the day of the trip.
7.8	7.8	Road leading to Greenville is on the left.
17.0	9.2	Outcrops of Seboomook Formation will be seen now for the next several miles (no stopping).
24.5	7.5	West Branch Penobscot River
26.2	1.7	Ragmuff Gate. This gate is locked and maintained by Great Northern Paper Company. For the rest of the trip we will be on roads to which there is no general access by the public. Speeds on these roads will be between 25 and 45 mph. Do not deviate from the route described or lag behind. The terrain for the next several miles is relatively flat. Outcrop is sparse. This is typical of terrain underlain by the Seboomook Formation.
32.8	6.6	Former site of Great Northern's Ragmuff logging camp.
40.6	7.8	Bear to left at Forestry sign.
40.9	0.3	Junction of Four Roads. Bear to right on the road which is maintained. (Not a sharp right onto the road with the grassy center leaving the gravel road on your left.)
41.2	0.3	Town line marker
44.3	3.1	Locked gate. We are now leaving Great Northern lands. Lands that we are now driving on are managed by Seven Islands Land Co. Continuing down this road, ridges which appear before us are the outcrop area of the Allagash Lake Formation.

46.3	2.0	Caucomgomoc Lake Dam. From the dam you have a panoramic view of Caucomgomoc Lake. Low hills to the north (right, as you face Caucomgomoc Lake), are the outcrop area of Allagash Lake Formation. The hills in the distance to the west (looking down the lake), is the outcrop area of the Frontenac. The lower hills, also to the west and closer to the lake, are Pre-Silurian outcrop. The trace of the Scott Brook fault trends approximately north-south between the dam and the Allagash Lake Formation. Scott Brook, Loon Stream and Ciss Stream all follow this fault. The Scott Brook fault separates the Allagash Lake Formation from the Seboomook Formation. Cross the dam and proceed.
47.9	1.6	LEFT TURN
50.9	3.0	Ciss Stream Bridge, cross and proceed.
51.5	0.6	BEAR LEFT and take the left arm of the "Y" at the intersection. Outcrops of the Allagash Lake Formation on either side of the road as you proceed.
53.2	1.7	Poland Pond. Brief stop.
53.3	0.1	STOP 1. Loon Stream Formation. This outcrop is typical of the greenish metapelite of this unit. Pervasive slaty cleavage strikes N 22° E with dips 47° to the northwest. Locally in this outcrop you may see kink bands, northwest trending fracture cleavage and two sets of steeply dipping quartz veins.
53.5	0.2	OPTIONAL STOP 1. Loon Stream Formation. This outcrop is less slaty in overall appearance. Metabasalt is present on the south side of the road. One may speculate that portions of this unit were originally a mafic or basaltic ash.
54.0	0.5	Loon Stream Formation (no stopping)
54.1	0.1	OPTIONAL STOP 2. Left turn. Proceed approximately 2.1 miles to the shore of Caucomgomoc Lake. At low water, a traverse of the shore line will exhibit in relative close proximity all of the major lithologies of the Caucomgomoc Lake Formation. Additionally, you may see many of the volcanic and sedimentary features of this unit.

54.2	0.1	STOP 2. Caucomgomoc Lake Formation. A large outcrop of metabasalt. This outcrop demonstrates textural variability of this unit on a small scale. Some units may be interpreted as volcanic ash while others may be interpreted as flows. Please be careful not to fall or dislodge rocks onto your colleagues below.
55.3	1.1	STOP 3. Caucomgomoc Lake Formation. Quartzose and feldspathic wackes. Lithologies and sedimentary features are fairly typical of the sandstones which are interbedded with the basalts of this unit.
57.9	2.6	LEFT TURN
62.4	4.5	Caucomgomoc Chain. This gate is operated by North Maine Woods.
63.2	0.8	STOP 4. Hurd Mountain Formation. Wackes and pelites of the undeformed lower portion of this formation.
63.5	0.3	OPTIONAL STOP 3. Left turn. Proceed down the road (but do not turn in to the camp yard on the left), as far as rationally possible, (approximately 0.5-0.75 mi). Park the vehicles and proceed to the lake shore. From there traverse east along the shoreline noting the basalts and sedimentary rocks of the Caucomgomoc Lake Formation.
63.8	0.3	Gate
64.2	0.4	RIGHT TURN
64.6	0.4	STOP and park on the side near the "Y" in the road. Here we will probably combine the party into the vehicles with highest clearance. We will proceed on the "right arm"of the "Y" (bear right). We will stop at several exposures of the Hurd Mountain Formation. This area exhibits the best outcrop and shows the variation in lithology and "tectonic style" of the melange. There have been several "wash outs" on the roads in this area. Drivers are urged to use caution and not tailgate. Return to vehicles and retrace route to Loon Lake road.

65.0	0.4	RIGHT TURN onto Loon Lake Road. Small outcrops of the Hurd Mountain Caucomgomoc Lake, and Loon Stream Formations are exposed as "pavement" along this road. No stopping.
68.7	3.7	OPTIONAL STOP 4 - LEFT TURN. This Woods road is passable with two wheel drive vehicles. Fallen trees may make passage difficult or impossible. You should be able to drive down this road for approximately two miles before it becomes impassable. Loon Stream Formation outcrops in the road as pavement sporadically. These outcrops provide you with typical lithology of the unit as well as structures commonly seen. From where the road becomes impassable, make your way to the shore of Caucomgomoc Lake and traverse northwesterly observing outcrops of Loon Stream Formation as you proceed.
72.6	3.9	LEFT TURN onto Scott Brook Road.
73.9	1.3	BEAR RIGHT at four road intersection. We will now retrace our route to Pray's store.
88.6	14.7	Ragmuff Gate. LEFT TURN.
107.0	18.4	RIGHT TURN for those of you who are going to GREENVILLE. Greenville is 35 miles. First gas station is in Kokadjo about 15-18 miles down this road. STRAIGHT for those of you who are going to Millinocket.
114.8	7.8	Pray's store.

TRIP D-3

MULTIPLE-TILL LOCALITIES IN CENTRAL MAINE

by

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INTRODUCTION

Purpose of Trip

The purpose of this trip is to introduce those interested in glacial deposits of New England to four remarkable multiple-till sites; one on Austin Stream, near Bingham and to three exposures along the Sandy River near New Sharon, Maine. There will also be an opportunity to consider the origin of multiple laminations of silt and clay interbedded with till at three of the sites.

Previous Work

Caldwell (1959, 1960) did some early work at the New Sharon sites. Caldwell (1983) concluded that an organic-rich sandy silt separated two tills, the lower till of Early Wisconsin age and the upper till of Middle and Late Wisconsin age. Borns and Calkin (1977) believed that there were three tills at New Sharon, representing Early, Middle and Late Wisconsin glaciation. Borns and Calkin also discussed the Austin Stream locality. Koteff (in press) believes there are Lower and Upper Wisconsin tills at the New Sharon, but no Middle Wisconsin till.

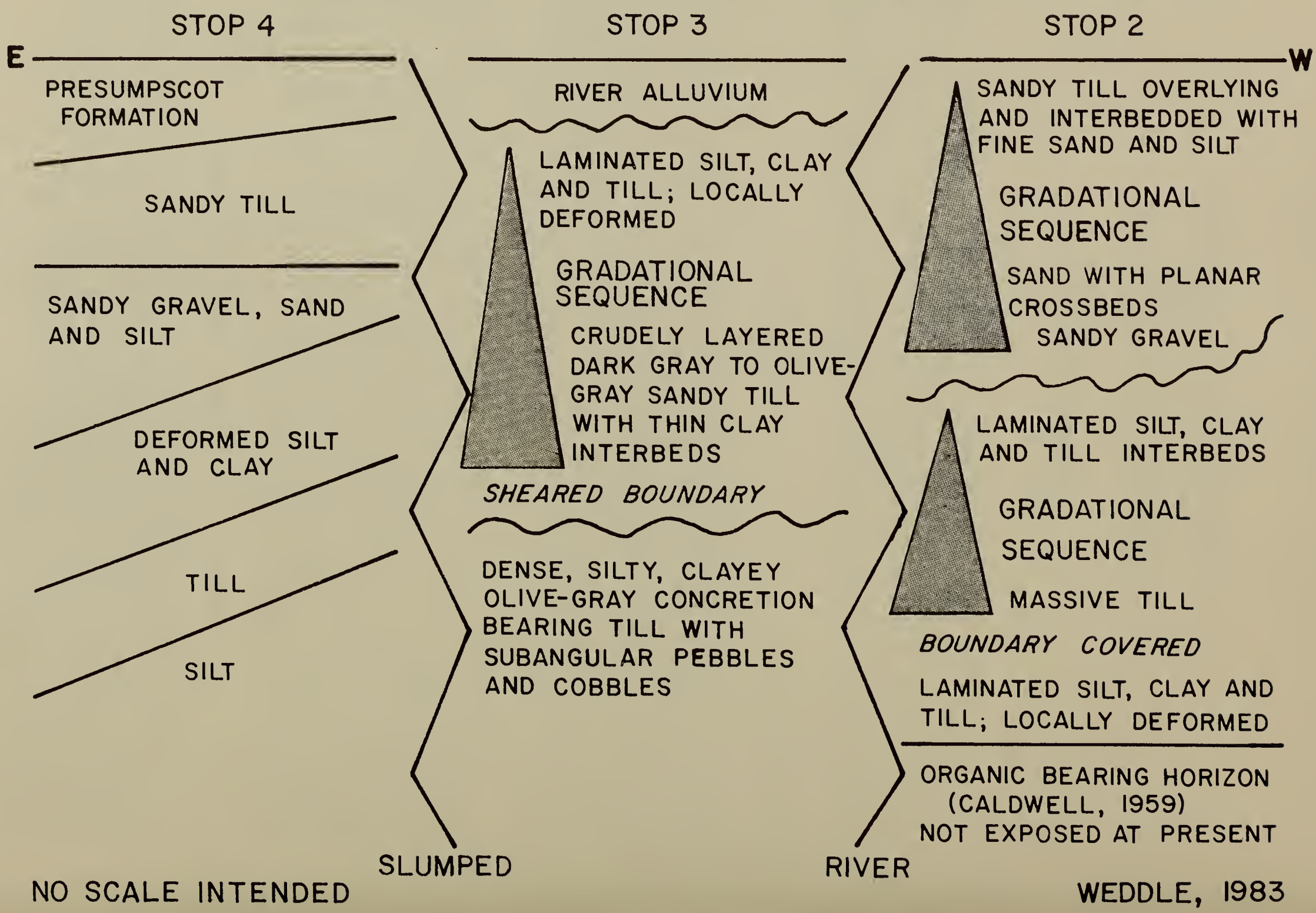
Previously unpublished pollen studies of the organic sediment by Caldwell and by Terasmae (personal communication, 1960) suggest that the climate during the deposition of the organic deposit was cooler than present, making the New Sharon organic deposit a possible equivalent of the St. Pierre beds of the St. Lawrence lowland. The pollen (see Table 1) also suggest that the organic sandy silt was deposited in a small lake or marsh. Numerous twigs and branches from the organic deposit were abraded like driftwood, further suggesting a lacustine environment of deposition.

CURRENT AND PROPOSED INVESTIGATIONS

Field studies conducted in the New Sharon gorge from mid-July to early September of 1983 have provided a basis for presenting generalized stratigraphic sections described herein. Three exposures occur along the Sandy River at New Sharon, Maine, east of where Route 2 crosses the river; one on the north bank and two on the south bank (from west to east, stopes 2, 3, and 4: Figure 1).

FIGURE 1.

SCHEMATIC STRATIGRAPHIC SECTIONS, NEW SHARON, MAINE



NO SCALE INTENDED

WEDDLE, 1983

TABLE 1. Pollen percentages in two samples of organic sandy silt, New Sharon

Pollen Type	Sample NS-4 (Caldwell)	Sample NL60 (Terasmae)
Tree Pollen		
Spruce	61.8	58.0
Pine	31.1	34.4
Fir	5.7	3.3
Hemlock	0.0	0.0
Alder	3.8	4.2
Other	--	--
Non-Tree Pollen		
Grass	0.6	1.6
Sedge	3.0	1.6
Waterlily	0.6	--
Spores		
Ferns	1.2	0.5
Equisetum	0.6	--
Sphagnum	1.3	3.1

During the summer, logging of sections along the exposed bluffs was conducted utilizing the method described in Eyles Eyles (1983) and Eyles, Eyles and Miall (1983). Samples for grain-size analyses and mineralogical analyes were collected. Sample collection for paleomagnetic analyses was attempted, however the collection technique proved unsuccessful. Pebble counts and till fabric analyses were performed on the concretion-bearing lower till, and a till-fabric analyses was performed on the diamict immediately above the lower till at stop 3. Hopefully, some of this data will be presented at the field trip if it has been reduced in time.

Future studies in the gorge will concentrate on microscopic examination of diamict fabric, especially in the laminated sediments, and a successful attempt at collecting samples of the laminated sediments for paleomagnetic analyses. Also, a more widespread investigation of till stratigraphy in west-central Maine will begin in the field season of 1984.

From observations made during a field trip to northern New Hampshire, July 1983, it is obvious that the stratigraphy exposed at New Sharon is important with regard to correlation with the regional till stratigraphy determined in southern New England. Participants of that trip included Weddle, W. Thompson, W. Holland, R. Dineen, C. Koteff, and R. Goldthwait.

A visit to the Nash Stream site (Pessl and Koteff, 1970) found the section, so well exposed in 1970, to be nearly indiscernible due to slump and overgrowth. However, nearby in Milan, New Hampshire, an exposed pit along Rt. 110 revealed a stratigraphy similar to that exposed at Nash Stream, although not as complete. The proximity of Nash Stream and the New Sharon multiple-till sites invites correlation of "lower- and upper-till" at both sites, with regard to the southern New England till stratigraphy (Newton, 1979). However, the stratigraphy at New Sharon and its local, as well as regional significance is not completely clear. Much of the thickness at stops 2 and 3 comprise laminated sediments that appear similar to units described in Evanson, Dreimanis, and Newsome (1979) It is not clear how much of the thick diamict described at stop 2 consists of sediment deposited by grounded ice, or was deposited as subaqueous flow deposits. The environment of deposition for the laminated sediments and some of the thick diamict is envisioned as one similar to that described by Miall (1983, Fig. 6), although not necessarily associated with a glaciomarine environment described by Miall.

Further investigation over the next several field seasons will attempt to clarify the regional significance of the New Sharon site with the southern New England till stratigraphy.

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ROAD LOG

This trip begins at the parking lot of the Squaw Mountain Lodge at 8:30 A.M.

The Bingham quadrangle, the Norridgewock quadrangle, and the Farmington 15' or the New Sharon 7½' quadrangles cover the areas of the field trip stops.

Mileage (mileage is only approximate)

0	Leave Squaw Mountain Lodge.
1.5	Turn right on Rt. 15.
7.5	In Greenville, turn right at blinking light, continue south on Rt. 15.
24.5	Monson slate quarries in Carrabassett formation provided material for telephone switchboards and the shingles on nearly all roofs in this part of Maine.
31	Titcomb's store, Abbot village. Turn right on Rt. 16. Titcomb's has the best ice cream in the State of Maine.
55	Turn right on wood hauling road, formerly roadbed of Bingham and Moosehad Lake Railroad.
60	STOP 1. Austin Stream, Moscow. Thick till sections on north bank of stream contain multiple laminations of silt and clay interbedded with till. Some laminations appear to be glacio-lacustine deposits while others are not graded and may be subglacial deposits.

Return to cars, turn around and return to Rt. 16.

65. Turn right onto Route 16.

66. Junction with US 201. Turn left.

67. Turn right on Route 16. Presumpscot formation underlies Emden formation on south side (left) just before bridge over Kennebec River.

67.5 Turn left (south) on 16.

69.0 Large esker that is part of system that stretches from Parlin Pond north of here to a glacio-marine delta in Augusta.

84. North Anson. Continue south of 201-A. Route 16 turns right.

89. Anson. Turn right of Route 43.

98. STOP 2. North Bank of Sandy River.
The approximate thickness of sediments exposed on the north bank is 30 meters (m). Four meters of laminated sediments are overlain by 20 m of olive- to dark-gray massive diamict (Munsell color 5Y 5/2; 4/1), which grades upwards transitionally into laminated sediments. This unit is overlain by a 6 m sequence of planar cross-bedded sand and sandy gravel that grades upwards into fine-grained sand and silt; the silt and sand are interlayered with and overlain by a sandy, compact diamict.

The laminated sediments exposed beneath the massive diamict consist of interbedded silt, clay, and diamict. Individual layers generally vary from less than 1 cm to 5 cm thick; some diamict layers are greater than 1 m thick. Silt and fine-grained sand intraclasts occur within some of the clay layers. Subangular pebbles, generally 1 to 3 cm in length, occur in the diamict; some boulders are greater than 0.5 m in width. Many pebbles and boulders appear to have been deposited as dropstones and deform underlying strata, and are draped by overlying strata. Flow folds and flow structures are rare in these sediments. Two lineation sets were observed on the bedding-plane surface of the fine-grained laminated sediments. One lineation set is apparently due to the intersection of vertical clay stringers (water-escape structures ?) with the bedding plane surface. The other lineation set may possibly represent striae due to shearing of the laminated sediments by overiding ice, or could represent sole mark striae due to subaqueous flow-diamict sliding over the finer grained unit. The laminated sedments

are locally contorted and deformed, especially at the west end of the bluff. Shear zones(?) and thrust faults also occur at this stop, but have not been examined in detail.

The two exposures along the south bank are separated by a long stretch of slumped and overgrown riverbank. Two macroscopically distinct units occur at Stop 3. A dense, silty, clayey olive-gray concretion-bearing till is overlain by a crudely layered, dark gray to olive-gray sandy diamict with clay interbeds. The overlying diamict appears to be in shear contact with the underlying till. Layers and stringers of the concretion-bearing lower till are incorporated into the overlying diamict; concretions in these till layers and stringers are nearly parallel to the layering. No concretions are found in the overlying diamict, however, as clasts.

The layering in the overlying diamict becomes more distinct transitionally up-section, where the unit is a finely laminated sequence of silt, clay, and diamict interbeds. Some of the diamict interbeds are several centimeters thick, and flow structures and syndepositional deformation structures are found in them. Minor normal and reverse faults also are present, and offset these units.

At the east end of this exposure, approximately 4.6 m of the lower till is overlain by 2 m of diamict and laminated sediments. At the west end of the exposure, approximately 6 m of diamict and laminated sediments occur. The contact between the lower till and overlying diamict is obsured by slumped riverbank for most of the exposure.

STOP 4 is the least studied section exposed along the Sandy River in New Sharon. Its stratigraphic relationship to the units exposed upstream is unclear. The 20 m of sediments exposed here are generally sandier than the units of Stops 3 and 2. These units consist of silt overlain by dense, silty, clayey diamict, overlain by deformed silt and clay interbeds. These units are overlain by locally deformed sand and sandy gravel, in turn overlain by a sandy, dense till. The Presumpscot Formation (?) occurs at the top of the sequence, overlying the till.

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